GEOMORPHIC AND STRATIGRAPHIC DEVELOPMENT OF LAKE BONNEVILLE'S INTERMEDIATE PALEOSHORELINES DURING THE LATE PLEISTOCENE

by

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STATEMENT OF DISSERTATION APPROVAL

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ABSTRACT

During the height of the Last Glacial Maximum (LGM), multiple paleoshorelines of late Pleistocene Lake Bonneville formed within the Bonneville basin of western Utah, eastern Nevada, and southeast Idaho. The geomorphic and sedimentological history of the paleoshorelines related to the relict lake has been used as proxies to understand paleoclimatic shifts during the LGM. The depositional and chronologic history of the lake's more significant paleoshorelines is well established. However, multiple transgressive paleoshorelines, termed "Intermediate" paleoshorelines, have also been identified but poorly documented in the basin. Intermediate paleoshorelines are found between the altitudinal limits of the lake's Bonneville and Provo levels. The variations in the altitudinal limits of these paleoshorelines make their correlation and chronologic record difficult to decipher.

Geologic maps were produced for unconsolidated sediments near Stockton, Utah and in the Hogup Bar located southeast of Park Valley, Utah, to provide a geologic framework of the stratigraphic and geomorphic developmental and chronological record related to the Intermediate paleoshorelines. New stratigraphic and chronological data provided from these maps record two previously unpublished oscillatory events as well as evidence for previously proposed oscillatory events that occurred during the Intermediate (transgressive) phase of the lake. A model was also developed to correlate the Intermediate paleoshorelines within the basin, by updating and incorporating data from

hydro isostatic rebound models, by incorporating data from a model that predicts potential wave energy, and by incorporating sedimentological and geomorphic data from the paleoshorelines to explain why variations exist in the altitudes of the Intermediate features. The potential correlation of six significant and multiple less substantial Intermediate paleoshorelines suggest that the chronologic record of these features can be established. However, more chronological and sedimentological evidence needs to be obtained before the proposed chronology of the Intermediate paleoshorelines and the observed oscillatory events can fully be demonstrated. As further documentation and chronology of these intermediate paleoshoreline features are obtained it will elucidate how the lake has responded to past submillennial climatic shifts associated with the LGM and will lead to a better understanding of the risk associated with future oscillations of the lake's surface.

This thesis is dedicated to the most important people in my life, my family. I dedicate this work to my wife, Traci, who has supported me in all of my pursuits, and to my children Avery, Kendric, Emmalee, and Caleb who have heard me say many times during the last 6 yrs “I need to work, can we do that later?” Thank you for your understanding and support.

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CHAPTER 1

INTRODUCTION

Objectives

This dissertation and the concepts discussed therein address the late Pleistocene deposits of Lake Bonneville and specifically a subset of transgressive deposits referred to as the Intermediate paleoshorelines. Lake Bonneville is a large lake that occupied the Bonneville basin of northwestern Utah, southern Idaho, and western Nevada during the late Pleistocene termination of the Last Glacial Maximum (LGM) (Fig. 1). Continual changes in the lake's water budget and resultant water level have provided a substantial repository of large sedimentological and geomorphic paleoshoreline features within its basin. The persistence of the lake's water level depended on either threshold controls as the lake flowed out of the basin or stabilizations of the regional/global climate. Studies related to these lacustrine deposits have provided researchers with a general trend of the basin's broad climatic history over the last 30,000 yrs (Fig. 2). Even though the general representation of the basin's lacustrine chronology is well documented, there are many smaller paleoshorelines that have not been documented and discussed. It is proposed that these smaller paleoshorelines record submillennial changes in the lake's hydrologic budget during the lake's transgressive and regressive stages. The evidence of the lake's temporary levels are closely tied to climatic responses; therefore, resolving its sedimentary development at a higher resolution has the potential to vastly improve the resolution of regional paleoclimate models and will aid in the exploration of how the Bonneville basin may respond to future climatic changes.

This dissertation focuses on a subset of these paleoshorelines that Gilbert (1890) referred to as “Intermediate shorelines” since they were landforms that recorded the ancient lake extent that formed during the transgressive rise of the lake during the intermediate altitudinal extent of Provo and Bonneville age landforms. These Intermediate features are of particular interest because they were deposited during the height of the LGM and record how Lake Bonneville responded to global and/or regional climatic drivers during the global event.

The objectives of this study include the following: 1) Collect a set of stratigraphic, geomorphic, and chronologic data by mapping areas that exhibit excellent examples of the Intermediate paleoshorelines. This first objective was met by mapping the late Pleistocene deposits related to the occupation of Lake Bonneville in the regions of northern Rush Valley near Stockton, Utah and the northern Hogup Mountains near Park Valley, Utah. 2) Use the data from the mapping projects to examine the broad lacustrine and climatic record of the lake and to determine how the Intermediate paleoshorelines relate to the current model of the lake’s history. 3) Determine what sedimentological and geomorphic factors affected how the Intermediate paleoshorelines formed and how these features were preserved in the lacustrine record. 4) Provide a correlation model that can be used to aid in the correlation of Intermediate paleoshorelines. 5) Explore the evidence for the hypothesis that Gilbert (1890) suggested regarding the Intermediate paleoshorelines as being evidence for an oscillating lake. 6) Discuss how the Intermediate paleoshorelines relate to regional and global climatic drivers.

Terminology

Table 1 consists of a summary of terms that will be useful in understanding this dissertation. The following descriptions are more detailed explanations of these terms and how they relate to one another.

The term shoreline as described by the U.S. Army Corps of Engineers (2003) is "the line of demarcation between a shore and the water or the intersection of a specified plane of water with the shore or beach." Past researchers of the Bonneville basin have used the term "shorelines" to define the landforms related to the ancient intersection of Lake Bonneville with the land surface. However, this term indicates an active body of water and is not a proper term for ancient coastal beaches where the water body is no longer present. Therefore, in this study the term paleoshoreline is used to delineate the physical linear expression of an ancient water body with the land.

Four large terminal lakes (lakes without an external outlet) have been identified in the Bonneville basin during the Quaternary (McCoy, 1987; Balch et al., 2005). Each of these distinct lakes is part of a separate lake cycle where Lake Bonneville is the most recent lake within these lake cycles (Fig. 3). Lake cycles are defined as distinct wetter periods in which a lake has occupied the basin separated by arid periods in which very small lake systems comparable to today's hydrologic system exist. Each of these lake cycles has a transgressive (a rising water level) and a regressive (a falling water level) phase of the lake cycle. Lake levels are a general term describing the mean altitude of the lake's water surface during the occupation of the lake. In these transgressive and regressive phases, the altitude of the lake's water level may oscillate. During these oscillations, the lake level may rise (transgress) or fall (regress); however, the general trend of the lake's water surface will either gradually rise (transgressive) or fall (regressive). Within the dissertation, oscillations are referred to as significant changes in the altitude of the lake's level (i.e., resultant lake level variations of 10–45 m) and hydrologic budget that correspond to submillennial patterns, whereas fluctuations are defined as small changes in the lake level (i.e., resultant lake level variations of <10 m) and water budget that are the result of seasonal or decadal patterns. In addition to the closed transgressive and regressive phases of the Lake Bonneville lake cycle, the basin also exhibited a period in which the lake was an open basin (a lake with an external outlet) that is being referred to as its open basin phase.

A few distinctive lake levels have been named and correlated throughout the Lake Bonneville lake cycle (e.g., Stansbury, Bonneville, Provo, or Gilbert levels). As in all lakes, deposits and landforms at a specific altitude do not represent the sedimentological record of a lake with a specific still water level (SWL). Deposition can occur throughout the shorezone especially in gravel beaches similar to the depositional environments of Lake Bonneville (Blair, 1999). The altitudinal crest of a gravel landform often does not mark the SWL but often marks the extent of wave run up during storms (Lorang, 2002; Buscombe and Masselink, 2006; Anthony, 2008; Carling et al., 2011). Therefore, when referring to a certain lake level (e.g., the Bonneville level), the altitude of that level/stage is referring to a mean altitude of the lake's SWL and the deposits or paleoshorelines related to the specific SWL exhibited within a relative range of altitudes around the mean. In addition to the four distinct lake levels (Stansbury, Bonneville, Provo, and Gilbert levels) in which the lake resides at or near a specific altitude, the lake has deposits associated with lake levels that developed during periods in which the lake did not reside for long durations. Since there are hundreds of paleoshorelines that can be mapped in the basin, these short lived paleoshorelines are categorized into groups of paleoshorelines that were either deposited in the relative rise (transgressive phase) or fall (regressive phase) of the lake's water surface. As the large hydrostatic load of the lake depresses the basin, the paleoshorelines will adjust to the increase of accommodation space; however, once the lake is gone the basin will rebound and cause differential altitudinal changes of these paleoshoreline features. The isostatically corrected altitudinal and age ranges of the paleoshorelines, associated with the more stable lake levels or associated with the short lived transgressive or regressive paleoshorelines, can be seen in Table 2 or seen schematically in Figure 2.

General Late Pleistocene History of Lake Bonneville

The pioneering studies of G. K. Gilbert (1890) initiated research regarding Lake Bonneville, and this terminal basin has continually been studied throughout the 20th century to understand a variety of geodynamic, geomorphic, paleontological, sedimentological, and anthropologic processes (e.g., Oviatt and Thompson, 2002). Gilbert (1890) described the geomorphic features of the Bonneville basin and established the relative timing of its various major paleoshorelines; however, the advent of radiometric dating techniques provided researchers with a much more robust chronologic record of the lake's major changes of water level and areal extent (e.g., Currey and Oviatt, 1985; Oviatt et al., 1992; Oviatt, 1997; Godsey et al., 2005).

The general hydrologic chronology and the broad climatic history inferred by the four major paleoshorelines of the lake cycle (i.e., Stansbury, Bonneville, Provo, and Gilbert lake levels; Fig. 2) have been well established (Oviatt, 1997; Kaufman, 2003; Balch, 2005; Godsey et al., 2005; 2011; Oviatt et al., 2005). Many less prominent paleoshoreline features formed during the transgressive and regressive history of the lake can also be found in the basin. Even though researchers have briefly discussed these less prominent features (Gilbert, 1890; Scott et al., 1983; Burr and Currey, 1988; Scott, 1988; Oviatt et al., 1994; Sack, 1999) and periodically included them in regional geologic maps (Miller & Oviatt, 1994; Miller & McCarthy, 2002), the literature does not describe their relevance in much detail.

Currey and Oviatt (1985) and Oviatt (1997) have suggested that multiple significant (scale of 20-45 m) oscillations occurred during the transgression of the lake. The Stansbury Oscillation(s) (~24,000–26,000 ^{14}C yr B.P.) may be composed of two separate oscillations and is the most prominent and well studied oscillation and may be composed of two separate oscillations (Oviatt et al., 1990; Patrickson et al., 2010). Oviatt (1997) suggested up to three more transgressive oscillations (termed U1–U3) from 17,000 to 24,000 ^{14}C yr B.P. The U1–U3 oscillations occurred during the period of the Intermediate paleoshorelines and have been tentatively correlated to global climatic events, such as Heinrich events and other 1,000 – 1,500 yr climatic cycles (Oviatt, 1997).

The Lake Bonneville lake rose to its maximum altitudinal limit (~1,552 meters above sea level (masl)) at ~15,500 ^{14}C yr B.P. (Oviatt, 1997) and roughly correlates to the timing of the LGM of glaciers within the Bonneville basin (Licciardi et al., 2004; Refsnider et al., 2008; Laabs et al., 2009; 2011). Once the lake level reached the altitude of a topographical threshold near Zenda, Idaho, the lake then overflowed into the Snake River/Columbia River Drainage Basin. When the lake reached this threshold, the water level within the lake stabilized for ~1,000 yrs to form the paleoshorelines of the Bonneville level (Oviatt et al., 1992; Godsey et al., 2005). Consistent hydro isostatic adjustment of the region depressed the basin floor during the duration of the lake at the Bonneville level and caused some Bonneville paleoshorelines to have multiple altitudinal expressions (Gilbert, 1890; Burr and Currey, 1988). The surface area and volume of the lake at the Bonneville stage has been calculated at ~51,556 km^2 and ~10,494 km^3, respectively. To aid in the understanding of the relative size and volume of the lake, the ancient lake is compared to the Great Lakes of the northeastern United States (Table 1.3). Lake Bonneville is comparable in surface area to the modern Lake Michigan but had a volume a little less than the modern Lake Superior.

The threshold at Zenda was composed of the weakly consolidated Salt Lake Formation and other unconsolidated alluvial deposits (Janecke and Oaks, 2011). Around ~14,500 ^{14}C yr B.P. (O'Conner, 1993) the lake catastrophically breached the threshold, and that breach resulted in a drop in its water level by ~108 m and a loss of ~5,238 km^3 of water. The volume of water lost during the flood can be compared to slightly more than the modern Lake Michigan and was ~ half of the volume of the lake (Table 1.3). The flood flowed into Marsh Valley, Idaho; the Snake River; the Columbia River; and then out to the Pacific Ocean. Following the flood, the lake was then constrained by the continued overflow of a new topographical divide south of Red Rock Pass, Idaho (Janecke and Oaks, 2011). Due to the lake overflowing at this new threshold, the water level remained relatively constant, and significant paleoshorelines then developed at the lake stage known as the Provo level until ~12,500 ^{14}C yr B.P. (Godsey et al., 2011). However, just like the Bonneville level, it has been suggested that the basin isostatically adjusted to the new volume of water at the Provo level, causing the expressions of the Provo paleoshorelines at a range of altitudinal limits (Burr and Currey, 1988; Godsey et al., 2005; 2011).

Following the occupation of the lake at the Provo level, the water level quickly fell below historic levels for the modern Great Salt Lake (Benson et al., 1992; 2011; Oviatt et al., 2005). This dramatic regression has been related to the Bøelling/Allreød interstadial (Benson et al., 2011; Godsey et al., 2011) and is thought to have lasted ~500–1,000 yrs. The lake then transgressed briefly to the Gilbert paleoshorelines (~ 10,000 ^{14}C yr B.P.) and may be related to the cold period of the Younger Dryas stadial (Oviatt et al., 2005).

Dissertation Layout and Contributions

The following section describes the layout of the dissertation and how the various chapters are planned or in the process of being published. Because each chapter is to be published as separate manuscripts there may be repetition in the general background information given within the chapters.

Chapter 2 is a discussion and geologic map (Plate 2.1) of the Quaternary geology of the southern portions of the South Mountain and the Stockton 7.5' quadrangles of Rush Valley, UT. An adapted version of Chapter 2 is scheduled to be published with the Utah Geological Survey as an Open File Report once the northern portions of the Stockton and South Mountain quadrangles are mapped. Rush Valley is a subbasin of the main Bonneville basin. This subbasin is separated from the main Bonneville basin by a threshold consisting of a series of sizeable Intermediate and Bonneville age spits and bars that Gilbert (1890) referred to as the "Great bar at Stockton." The altitudinal limit of the threshold into Rush Valley is well above the Provo level; therefore, it is hypothesized that the paleoshorelines in Rush Valley should be represented by a series of Intermediate and Bonneville age paleoshoreline features. However, Burr and Currey (1988) suggested an alternative hypothesis for the paleoshorelines that are within Rush Valley. Burr and Currey (1988) suggest that paleoshorelines below the Bonneville level are not Intermediate in age, but remnants of smaller lakes that were impounded in the subbasin following the regression caused by the Bonneville flood. Chapter 2 describes the paleoshorelines in the Rush Valley and then discusses the geomorphic and stratigraphic evidence for the chronology of these paleoshorelines. The chapter then discusses the evidence for and against the two competing hypotheses for the Rush Valley paleoshorelines and how these paleoshorelines fit in with the understanding of the main Bonneville basin.

An adapted version of Chapter 3 is currently in the review process with the Utah Geological Survey and will be published as an Open File Report (scheduled 2012).

Chapter 3 is a discussion and geologic map (Plate 3.1) of the Quaternary geology of the Hogup Bar 7.5' quadrangle, in northwestern Utah. The area has well preserved paleoshoreline deposits and erosional features that were formed during the occupation of all four of the lake's major levels and numerous other less substantial lake levels, including the record of the Intermediate paleoshorelines. The description and map of the area included in this chapter provide a basis for investigating the preservation, development, and relationship of the major paleoshorelines with the Intermediate paleoshorelines.

In his 1890 Monograph regarding Lake Bonneville found it difficult to correlate the Intermediate paleoshorelines due to the inconsistency of the individual altitudes of the features and the inconsistency of the preservation of the number of these paleoshorelines at individual localities. Chapter 4 discusses the sedimentological, geophysical, and geomorphic factors that influence how the Intermediate paleoshoreline formed, why the features are preserved in the geologic record, and why the altitudinal crests of the individual features vary. The altitudinal variation of these paleoshorelines is quantified at multiple test localities and six (6) significant Intermediate paleoshorelines, and multiple smaller paleoshorelines, are correlated at these test localities. These paleoshorelines hypothesized to also be correlated throughout the basin. An accurate chronology and correlation of these Intermediate paleoshorelines will aid in the understanding of how the lake responded to regional and global climatic events on a submillennial scale.

Gilbert's (1890) hypothesized that altitudinal variations exhibited by the Intermediate paleoshorelines are a result the autogenic sedimentary processes of an oscillating lake. Chapter 5 describes the stratigraphic evidence for multiple oscillatory events in the Hogup Mountains and relates these events to other known in the basin. The chapter suggests that at least three relatively large oscillations (25–45 m), two that were previously undocumented, can be inferred in the Hogup Mountains. Combining these oscillatory events with other hypothesized oscillations suggests that there are six to seven (6 -7) proposed oscillations during the transgressive phase of the lake. However, with the relative low number of radiometric and sedimentological evidence around the basin the uncertainty for the actual number, amplitude, and timing of these oscillatory events are still relatively high. Therefore, until these details are better resolved, it will be difficult to understand how the pattern of these oscillatory events relate to global and regional climatic patterns.

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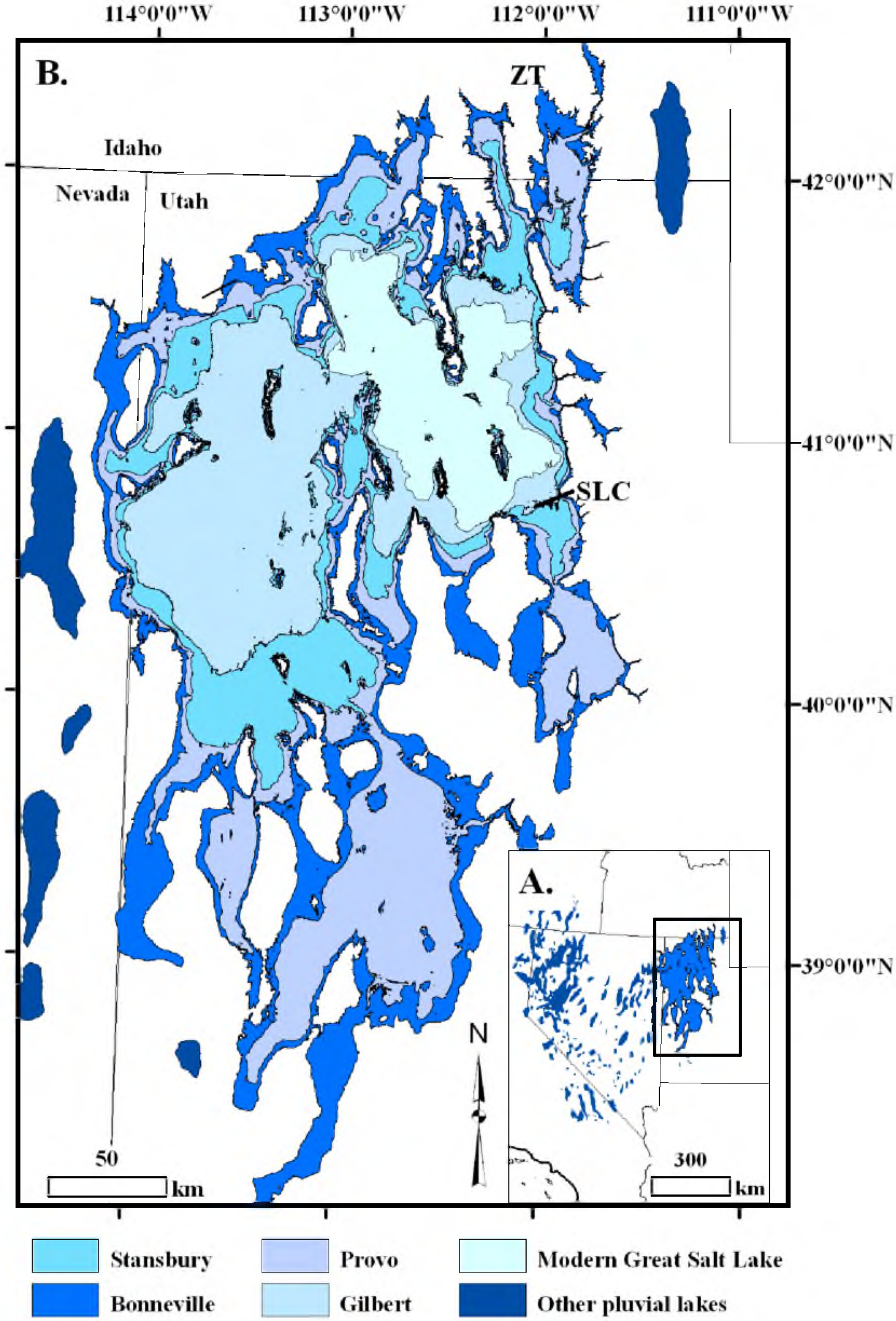
Figure 1.1: Lake Bonneville maximum extent of major lake levels. A) Regional map of the extent of Lake Bonneville at its maximum in relation to the maximum of other pluvial lakes in the Great Basin; B) The maximum extent of the modern Great Salt Lake and the maximum extent of each of the major lake levels of the Lake Bonneville lake cycle.



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Other pluvial lakes

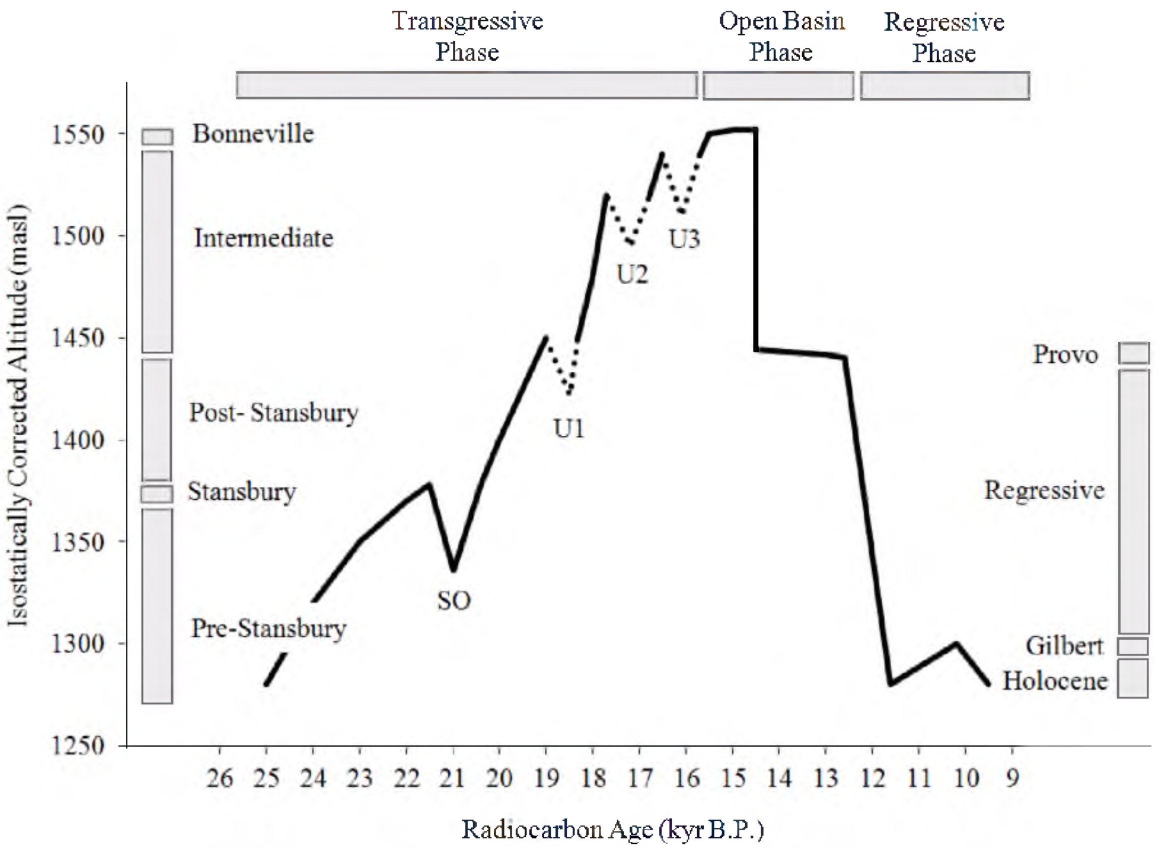

Figure 1.2: Lake Bonneville hydrograph modified from Oviatt (1997) and Godsey et al. (2011). Altitudes are adjusted for effects of differential isostatic rebound in the basin (Oviatt et al., 1992). Amplitude limits of lake stage fluctuations associated with the U1, U2, and U3 oscillations are approximate and are shown here schematically. The temporal range of the transgressive, regressive, and open phases of the lake cycle are shown horizontally, whereas the altitudinal range of each of the paleoshoreline groups are shown vertically within either the transgressive or the regressive phases of the lake cycle.

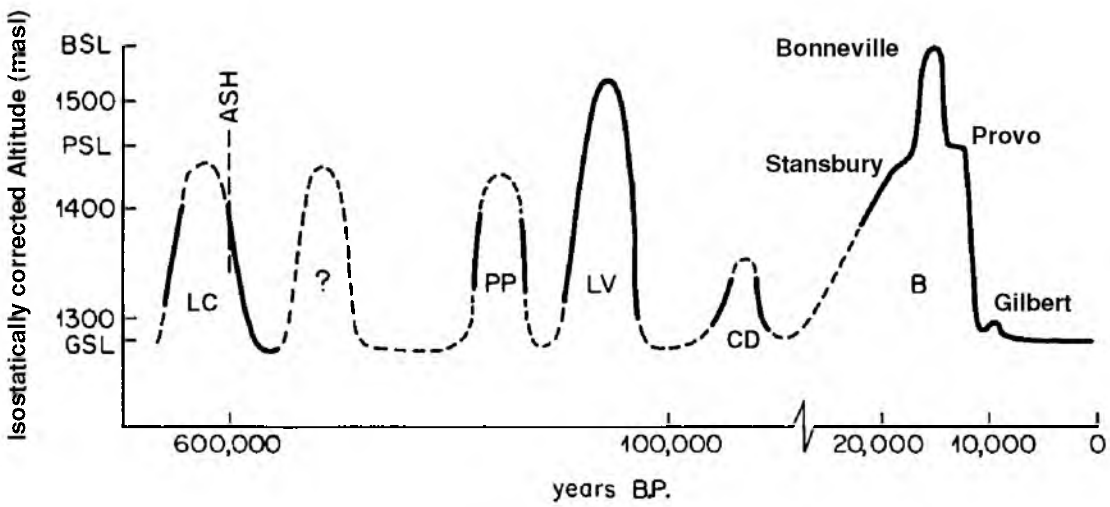

Figure 1.3: A schematic hydrograph of lake cycles in the Bonneville basin in the past 700,000 yrs (modified from McCoy, 1987). Abbreviations are as follows: B (Bonneville), CD (Cutler Dam), LV (Little Valley), PP (Pokes Point), and LC (Lava Creek) lake cycles. Isostatically corrected altitude in meters above sea level (masl).

Table 1.1: Common terminology used in the dissertation.

Basic terminology related to shorelines and paleoshorelines	
Paleoshoreline:	physical and geomorphic evidence of the shoreline of relict water bodies (e.g., lakes, oceans). In the context of the dissertation, it is the relict shoreline expressions of Lake Bonneville (Atwood, 2006).
Shoreline:	line of demarcation between a shore and the water or the intersection of a specified plane of water with the shore or beach (U.S.A.C.E., 2003).
Basic terminology related to the Lake Bonneville lake cycle	
Lake cycle:	complete rise and fall of a lake within a basin. The duration of the lake cycle is demarcated by the period in which the specific lake had lake levels above the modern altitudes of lakes within the basin. The lake level maximums record periods of wetter and/or colder climates; whereas playas develop during periods of arid climates.
Lake level:	general (mean) altitude of the lake's water surface during a defined period of the occupation of the lake in the basin (i.e., Stansbury, Bonneville, Provo, and Gilbert levels).
Lake level fluctuations:	small changes in the altitude of the lake's water surface (i.e., lake level variations of <10 m) and resultant water budget that correspond to decadal or seasonal patterns.
Lake level oscillations:	Significant changes in the altitude of the lake's water surface (i.e., lake level variations of 20–45 m) and resultant water budget that correspond to millennial or centennial patterns.
Open basin phase:	period in the lake cycle in which the water surface of the lake was relatively stable due to the lake overflowing into another basin.

Table 1.1: *(continued)*

Regressive phase:	period in the lake cycle in which the water surface of the closed basin lake was generally falling.
Transgressive phase:	period in the lake cycle in which the water surface of the closed basin lake was generally rising.

Groups of Paleoshorelines of the Lake Bonneville lake cycle (in chronologic order) – See Figure 1.2 or Table 1.2 for more detail.

Pre Stansbury paleoshorelines:	a series of paleoshorelines that developed during the rise of the lake (transgressive phase) prior to deposition of paleoshorelines related to the Stansbury lake level. The paleoshorelines lie between the altitudinal range of the modern lake level of the Great Salt Lake and the Stansbury lake level.
Stansbury paleoshorelines:	a series of paleoshorelines (transgressive phase) that developed when the lake level oscillated near the Stansbury lake level.
Post Stansbury paleoshorelines:	a series of paleoshorelines that developed during the rise of the lake (transgressive phase) following the deposition of paleoshorelines related to the Stansbury lake level. The paleoshorelines lie between the altitudinal range of the Stansbury and Provo lake levels.
Intermediate paleoshorelines:	a series of paleoshorelines that developed during the rise of the lake (transgressive phase) following the deposition of the Post Stansbury paleoshorelines. The paleoshorelines lie between the altitudinal range of the Provo and Bonneville lake levels.
Bonneville paleoshorelines:	a series of paleoshorelines (open basin phase) that developed during the maximum extent/level of the Lake Bonneville lake cycle. The lake level was relatively stable during the Bonneville level due to the lake overflowing a natural threshold into the Columbia River Drainage.

Table 1.1: *(continued)*

Provo paleoshorelines:	a series of paleoshorelines (open basin phase) that developed near the Provo level following the regression caused by the Bonneville flood. The Provo level was relatively stable due to the lake overflowing a bedrock threshold into the Columbia River Drainage.
Regressive paleoshorelines:	a series of paleoshorelines that developed during the fall of the lake (regressive phase) following the deposition of paleoshorelines related to the Provo level. The paleoshorelines lie between the altitudinal range of the Provo and Gilbert levels.
Gilbert paleoshorelines:	a series of paleoshorelines (regressive phase) that developed when the lake level oscillated near the Gilbert level.

Table 1.2: Age ranges and general altitudes of Lake Bonneville paleoshorelines within the basin.

Lake Cycle and Phase	Paleoshoreline	^{14}C yr B.P.	Cal. Yr B.P.[1]	Isostatically Corrected Altitude[2] (meters)
Lake Bonneville				
Transgressive	Pre-Stansbury	~25,000-22,000[3]	29,000-23,200	~1320-1336
	Stansbury	22,000 – 21,000[4]	24,000-23,200	1339-1378
	Post-Stansbury	21,000~19,000	23,200-19,200	~1378-1443
	Intermediate[6]	~19,000-15,300	20,400-18,300	~1449-1545
	Bonneville	15,300 – 14,500[7]	18,300-17,400	1545-1,552
Regressive	Provo	14,500 – 12,600[8]	17,400-15,000	1424-1445
	Regressive[9]	~12,600 – 11,000	15,800-10,800	~1295-1424
	Gilbert	11,000 – 10,000	10,800-9,300	1291-1296

[1]Calendar-calibrated ages of most paleoshorelines have not been published. Calendar-calibrated ages shown here have been estimated by using Calib 6.0 (Stuvier & Reimer, 1993).

[2]Paleoshoreline elevations were corrected by using the methodology of Currey and Oviatt (1985).

[3]Estimated based on altitude in comparison to the lakes hydrograph published by Oviatt (1997). – Dates have not been calibrated past 25,000 calendar yrs B.P.

[4]Oviatt et al. (1990). Currey (written communication to the Utah Geological Survey, 1996) assumed a maximum age for the Stansbury paleoshoreline of 21,000 ^{14}C yr B.P., which is used in the conversion to calendar yrs.

[5]Lowr intermediate paleoshorelines are transgressive paleoshoreline features positioned in between the altitudinal limits of the Provo and Stansbury paleoshorelines. The extent of the age of the paleoshoreline features is based on the extent of radiocarbon dates summarized in Sack (1999) and Oviatt (1990).

[6]Upper intermediate paleoshorelines are transgressive paleoshoreline features positioned in between the altitudinal limits of the Provo and Bonneville paleoshorelines. The extent of the age of the paleoshoreline features is based on the extent of radiocarbon dates summarized in Sack (1999) and Oviatt (1990).

[7]Oviatt et al., (1992), Oviatt (1997)

[8]Godsey et al., (2011) revised the timing of the occupation of the Provo paleoshoreline and subsequent regression.

[9]Regressive paleoshorelines positioned in between the Gilbert and Provo levels of the lake system. The estimated age and altitude range of these paleoshorelines are based on the constraints of the Provo regression and the altitudinal limits of the Gilbert paleoshorelines.

Table 1.3: Isostatically corrected Lake Bonneville values.

Isostatically corrected calculations (updated methodology of Wambeam (2001) based on DEM's with a 5 m resolution) of the volume, area, and depth of Lake Bonneville in comparison with the modern Great Lakes of the northeastern United States. Data for the Great Lakes was acquired from EPA Atlas (2012). Altitude of the mean lake level is measured in meters above sea level (masl).

Lake	Surface Area (km^2)	Volume (km^3)	Altitude (masl)	Max Depth (m)
Modern Great Lakes				
Superior	82,000	12,000	180	407
Huron	60,000	3,500	176	228
Michigan	58,000	4,900	176	282
Erie	25,700	480	174	64
Ontario	19,000	1,640	75	245
Pleistocene Lake Bonneville				
Bonneville	51,556	10,494	1,552	352
Provo	38,369	5,256	1,444	244
Difference of Lake Bonneville and Provo following the Bonneville Flood	13,187	5,238	108	108

CHAPTER 2

GEOLOGIC MAP OF UNCONSOLIDATED DEPOSITS IN SOUTHERN PORTIONS OF THE SOUTH MOUNTAIN AND STOCKTON QUADRANGLES, TOOELE COUNTY, UTAH

Location and Geographic Setting

The Stockton and South Mountain quadrangles are located within the Bonneville basin, a subbasin of the larger Great Basin of the western United States (Plate 2.1). The quadrangles are located in Tooele County, Utah, and are positioned between Rush and Tooele Valleys. The study area for this map consists of the southern portions of both the Stockton and South Mountain 7.5 minute quadrangles. The study area is bounded on the east by the Oquirrh Mountains, the west by the Stansbury Mountains, and the north by South Mountain (Fig. 2.1). The city of Stockton, Utah lies just north of the study area, within the Stockton quadrangle, and can be accessed via State Highway 36, ~4.5 miles south of the city of Tooele, UT.

Historically, the mapped regions, and particularly the area of the Stockton quadrangle, played an important role in the development of the mining industry in Utah. The Stockton Mining District was an excellent source of lead, zinc, silver, gold, and copper during the late 1800s (James and Atkinson, 2006). The district went through periods of mining activity and inactivity up to the mid 1960s.

Rush Valley is a terminal, structural subbasin of the larger Bonneville basin. Rush Lake is principally fed by groundwater sources and is the main hydrologic feature within the valley. Other hydrologic features located in the study area include a series of ephemeral drainages: Soldier Canyon, which originates from the Oquirrh Mountains, and Welsh and East Hickman Canyon, which originate in the Stansbury Mountains.

General Quaternary Geology

Debris flows and sheetwash originating from the Oquirrh and Stansbury ranges formed substantial alluvial fans in Rush valley. During the late Pleistocene, a large pluvial lake known as Lake Bonneville occupied the northern portion of Rush Valley. The initial rise of the Lake Bonneville cycle occurred ~ 27,000 ^{14}C yr B.P., reached its maximum extent ~ 15,500 ^{14}C yr B.P. near the end of the Last Glacial Maximum, and regressed to the modern elevations of the Great Salt Lake by ~ 10,000 ^{14}C yr B.P. (Oviatt, 1997). Four significant paleoshorelines developed during the Lake Bonneville lake cycle, the Stansbury, Bonneville, Provo, and Gilbert levels, oldest to youngest, respectively (Fig. 2.2). The lake at its maximum extent, the Bonneville level, encompassed 51,556 km^2 of northwestern Utah, portions of southern Idaho, and eastern Nevada. This calculation was updated from an older calculation by Wambeam (2001) by using a more accurate 5 m isostatically corrected digital elevation model.

Within the main Bonneville basin, the lake reached the Stansbury level (~1350 masl) by ~ 20,000 ^{14}C yr B.P. (Oviatt et al., 1990). Since the lowest altitude for Rush Valley is well above the altitude range of the Stansbury level paleoshorelines, we can assume that if a lake system was present in Rush Valley, during the age of the Stansbury level, it must have been a separate small closed lake or open lake system controlled by the unknown altitude of the threshold between Rush and Tooele Valley.

Gilbert (1890) and Gilluly (1929) proposed that prior to the transgression of Lake Bonneville into Rush Valley, that Rush Valley was likely hydrologically connected to Tooele Valley and the main Bonneville basin. Both Gilbert (1890) and Gilluly (1929) suggested that the northern end of Rush Lake exhibits a channel like form. The alluvial channels south of Indian Hill (T. 5 S., R. 5 W., Sec. 16 and 17, Salt Lake Base Line and Meridian) and in Soldier Canyon (T. 4 S, R. 4 W., Sec. 31) are both oversized for the current ephemeral streams present in the channels. Our interpretation is that these oversized channels held considerably higher base flows prior to the occupation of the valley by Lake Bonneville than today, and that these streams may have flowed from Rush Valley into Tooele Valley near the saddle between South Mountain and the Oquirrh range (T. 5 S., R. 5 W., Sec. 23). As the climate changed and water levels of Lake Bonneville rose, the lake eventually advanced into Rush Valley sometime following 18,000 ^{14}C B.P. (Burr and Currey, 1988). As the water level of the lake rose, longshore currents transported sediments from substantial Tertiary alluvial fans on both the west and eastern sides of Tooele Valley to the south. These sediments accumulated across this saddle between South Mountain and the Oquirrh Mountains to form a sizeable baymouth bar and spit complex, known as the Stockton Bar. This bar continued to be built up as the lake level reached its maximum vertical extent at the Bonneville level. The altitude of the bedrock threshold beneath the bar is unknown; however, GPR profiles of the bar

(Smith et al., 2003) suggest that sediments are thicker near South Mountain and that the tributary would have been near the west side of the bar.

Gilbert (1980) named a series of transgressive paleoshorelines associated with transgressive phase of the lake as the “Intermediate” paleoshorelines. Intermediate paleoshorelines are transgressive deposits located between the altitudinal limits of the Bonneville and Provo paleoshorelines. At the Stockton Bar there are a series of prograding, Intermediate beach ridges (a-j) just north and south of the main Stockton Bar (Fig. 2.3), oldest to youngest, respectively. Intermediate beach ridges exhibit a northerly concave curvature suggesting that wave action that built these barriers originated from the north (Fig. 2.3). Following the regression of the lake, this baymouth bar acted as a sedimentological and hydrologic threshold separating the surface water sources of Rush Valley from Tooele Valley.

The lake stayed at the Bonneville level for roughly 1,000 yrs; however, the lake level catastrophically fell over 100 m when the lake breached a natural dam ~14,500 ^{14}C yr B.P. at a threshold near Zenda, Idaho, and flooded into the Snake River basin (Oviatt, 1997; O’Conner, 1993). Consequently, this very quick, regression separated Rush Valley from the main body of the Bonneville basin and created a lake system within Rush Valley that is hydrologically disconnected from the main body of the lake. Following this regression, Lake Bonneville’s lake level then stabilized at the Provo Level for ~2,000 yrs before a rapid, climatically induced regression ~12,600 ^{14}C yr B.P. (Godsey et al., 2011). Following this rapid regression, there was a brief oscillation at ~10,000 ^{14}C yr B.P, which is referred to as the Gilbert Level (Oviatt et al., 2005).

Late Pleistocene and Holocene Paleoshoreline Features

Within the study area (Plate 2.1), depositional and erosional traces of three significant paleoshorelines and multiple smaller paleoshorelines have been delineated. Of the four major levels of Lake Bonneville (i.e., Bonneville, Provo, Stansbury, and Gilbert), only paleoshorelines of the Bonneville level lake are present in the study area. The highest of the paleoshorelines are Bonneville level and can be distinctly identified as a result of wave processes within the lake eroded and reworked the deposits of the Soldier Fan (T. 5 S., R. 5 W., S. 5, 11, and 12) and Indian Hill (T. 5 S., R. 5. W., Sec. 8).

Two other significant paleoshorelines in Rush Valley termed by Burr and Currey (1988) as the Shambip (Z) and Smelter (Y) paleoshorelines lie in the altitudinal range of the Intermediate paleoshorelines (Table 2.1). Since Rush Valley was separated by the main basin following the Bonneville Flood, Burr and Currey (1988) suggested that the Shambip and Smelter paleoshorelines were not transgressive deposits, but regressive deposits related to Provo and Gilbert age deposits present in the main basin.

Bonneville Paleoshorelines

Bonneville age paleoshorelines developed during the occupation of Lake Bonneville at its maximum. Lake Bonneville landforms consist of gravel and sand (Qlgb) deposits from reworked alluvial fans (i.e., Soldier Fan and near Welch Canyon) and many erosional notches and strandlines on the west side of the Stansbury Mountains (see Table 2.1). The Bonneville paleoshoreline is the most recognizable paleoshoreline due to its distinct contrast with non lacustrine deposits and the stabilization of the lake's water level for ~ 1,000 yrs (15,300 – 14,500 ^{14}C yr B.P.: Benson and others., 2011;

Oviatt, 1997; O'Conner, 1993). During the ~ 1,000 yrs of Lake Bonneville, water overflowed the basin's former topographic threshold near Zenda, Idaho. The threshold is thought to have consisted of a substantial alluvial fan overlying the semi consolidated bedrock of the Tertiary Salt Lake Formation (Janecke and Oaks, 2011). The lake slowly cut into this threshold as the out flowing river flowed into the Snake River / Columbia River basin. The threshold catastrophically failed at ~ 14,500 ^{14}C yr B.P., causing the rapid fall of the lake level by ~ 108 m.

Shambip and Smelter Paleoshorelines

Shambip (Z) and Smelter (Y) paleoshorelines are below the Bonneville paleoshorelines and above the Holocene Rush Valley paleoshorelines (see Table 2.1). These features are exhibited in the study area as both erosional wave cut notches and depositional gravel (Qlg) bars and beach ridges. Since the altitude range of the deposits associated with both Provo and Gilbert lake levels are well below the altitude (1510 masl) of the basin floor of Rush Valley, neither of these paleoshorelines should be evident within Rush Valley. However, Burr (1989) and Burr and Currey (1988, 1992) hypothesized that the Shambip and Smelter paleoshorelines are not Intermediate paleoshorelines but regressive paleoshorelines formed when water was impounded in Rush Valley following the Bonneville flood. Burr and Currey (1988) suggest that the Shambip paleoshoreline may be equivalent with the timing of the Provo age lake, whereas the Smelter paleoshoreline may be equivalent to the Gilbert age lake. Lake Shambip is named after an Indian name given to the valley, where "Shambip" means Rush (James and Atkinson, 2006); Lake Smelter is named after an old smelter of the

Chicago Works facility that lies on one of the proposed Smelter paleoshorelines. Northerly trending waves from Rush Valley shaped erosional and depositional features related to these paleoshorelines at the south end of the Stockton Bar (T. 4 S., R. 5 W, Sec. 23). These erosional features cut into transgressive bars that exhibit northerly facing concave curvature and were deposited from longshore currents originated from Tooele Valley. Assuming that the Shambip and Smelter features at this locality were deposited in an isolated lake within Rush Valley, Burr and Currey (1988) interpreted the deposits to be regressive, and that they were related to the Provo and Gilbert age lakes. However, Burr and Currey (1988, 1992) did not have radiometric dates or stratigraphic evidence to support this hypothesis.

An alternative hypothesis for these two significant paleoshorelines is that both Lake Shambip and Smelter paleoshorelines were Intermediate paleoshorelines. The erosional processes that formed these paleoshorelines cut into the older transgressive beach ridges (a-f). The lateral shape of the deposits from the Smelter and Shambip paleoshoreline on the southern side South Mountain have a southerly facing concave shape. Erosional remnants of these paleoshorelines were most likely developed by waves that originated from the south. Therefore, these upper Rush Valley paleoshorelines were the remnants of the occupation of a water body in the basin during the transgressive rise of the lake. One of the main sources of evidence that these paleoshorelines were transgressive paleoshorelines was their correlation with other known intermediate paleoshoreline features in the main body of the lake basin. Heights of the crests (1540 m) of significant gravel deposits (bars) associated with the Shambip paleoshoreline near

Indian Hill, the Stockton Bar, and in the southeastern corner of the South Mountain Quadrangle, correspond to crest heights and sizes of similar intermediate gravel deposits on the northern side of South Mountain and the Stansbury Mountains. Similar significant paleoshoreline deposits near the altitudinal limits of the Rush Valley paleoshorelines can occur in other localities around the basin (i.e., Hogup Mountains, North Oquirrh Mountains). However, the uncertainty of the isostatic rebound of the basin, the different wave environments, and the various depositional and erosional conditions make it difficult to correlate paleoshoreline features based on altitude alone. In addition, due to the numerous Intermediate paleoshorelines in the Bonneville basin, the fact that these Intermediate paleoshorelines on the north side of South Mountain and other localities in the basin correlate to the same altitudinal limits of the Shambip and Smelter paleoshorelines could just be a coincidence.

To test these two hypotheses, the paleoshorelines were mapped within the study area and their stratigraphic and geomorphic relationships with known transgressive deposits from multiple localities were analyzed. In addition, where possible, material for radiometric dating was collected and analyzed. Near Indian Hill (T. 5 S., R. 5. W., Sec. 8) freshwater gastropod shells, Beta-307252 (*Stagnicola bonnevillensis*) and Beta-307253 (*Valvata utahensis*) provide a chronological constraint on the ages of the Shambip paleoshorelines. The age of the samples was 13,300 & 13,360 ^{14}C B.P., respectively. The stratigraphic interpretation of the sediment in which the samples were collected was over wash gravels (gravels deposited/washed over the barrier during a storm) behind a small beach barrier (Fig. 2.4). On the south side of South Mountain (T. 4 S., R. 5 W., Sec. 33) freshwater gastropod shells, Beta-307254 (*Valvata utahensis*), were collected and analyzed to constrain the age of the Shambip paleoshoreline. The age of the sample was 14,290 ^{14}C B.P. The stratigraphic interpretation of the sediment in which the samples were collected was offshore beach sands just below the Shambip paleoshoreline (Fig. 2.4). At both of these localities, no deepwater marl was found stratigraphically above these paleoshorelines and could indicate that the deposition was following the Bonneville age of the lake.

Based on these ages, the Shambip (Z) paleoshoreline was most likely deposited during the regressive Provo phase of the lake. The age of Smelter (Y) paleoshoreline is not constrained by radiometric ages due to a lack of suitable materials in sufficient stratigraphic relationships. Hence, it is not certain whether deposition occurred during the transgressive or regressive phase of the lake.

Rush Lake Paleoshorelines

Rush Lake paleoshorelines developed during the Holocene following the regression of the lake from the Bonneville level. See Table 2.1.

Quaternary Fault Scarps

A series of escarpments and lineaments in the alluvial fans of the western side of Rush Valley have been interpreted as fault scarps. The age of these small fault scarps/lineaments and the last rupture of these faults are uncertain. The faults are assumed to be normal faults related to the extensional forces of the Basin and Range tectonic region. Most of fault planes on the western side of Rush Valley are dipping to the east. However, there is one fault just east of Mormon Trail Road where the fault plane is dipping to the west and delineates the eastern side of a small graben.

Description of Map Units

Alluvial Deposits

Qal – alluvium (Holocene). Moderately sorted, interbedded gravels to cobbles within a matrix of sands and silts deposited in active alluvial channels, floodplains, and minor terraces. Coarser deposits found in proximity to mountain fronts and fining of deposits with distance from mountain fronts. The majority of these deposits are emitting from the larger drainages of Soldier Canyon, Welsh Canyon, Cow Hollow, and East Hickman Canyon. Other smaller gullies also have this type of deposit but are not significant enough to be mapped at this scale and are usually grouped with Qaf deposits. Maximum thickness is expected to be less than 20 feet (6 m).

Qaf_1 – level 1 alluvial fan deposits (Holocene to late Pleistocene). Post Bonneville fans that are poorly to moderately sorted, sub angular to rounded gravels to cobbles in a matrix of sand, silt, and clay. Fan deposits below the altitude of the Bonneville paleoshorelines usually consist of reworked Lake Bonneville gravels, marls (Qlm), and sands deposited (Qls) by ephemeral flooding (debris flows) and sheet wash events. Alluvial fan deposits originating from above the altitude of the Bonneville paleoshorelines tend to be coarse (cobble [64-256 mm]) and angular due to the erosion of the clasts from the proximal bedrock sources. In addition, deposits tend to be much coarser (cobbles) in the headland portion of the fans and become thinner and finer grained (coarse sand – pebble [0.5-16 mm]) in the more distal portion of the fan slope.

Qaf_1 deposits are equivalent to younger portions of Qaf_y deposits, but were delineated as Qaf_1 because younger fans or stream incisions do not cut them. Maximum thicknesses are expected to be less than 40 feet (12 m).

Qaf_y – young alluvial fan deposits (Holocene to late Pleistocene). Post Bonneville or Bonneville age fans that are poorly to moderately sorted, sub angular to rounded gravels to cobbles in a matrix of sand, silt, and clay. Qaf_y deposits are equivalent to Qaf_1 and Qaf_2 deposits; however, they are undivided because the specific age of the deposits cannot be determined. These deposits are above the elevation of known Bonneville age units and it is unknown if they were active during the occupation of the lake in the basin. Due to the proximity of local bedrock sources, the fan deposits tend to be coarser (cobble) and more angular than clasts within fans below the Bonneville paleoshoreline. Deposits tend to be much deposits tend to be much coarser (cobbles) in the headland portion of the fans and become thinner and finer grained (coarse sand – pebble) in the more distal portion of the fan slope. Maximum thicknesses of the deposits are expected to be less than 40 feet (12 m).

Qaf_2 – level 2 alluvial fan deposits, Bonneville lake cycle (upper Pleistocene). Poorly to moderately sorted, subangular to rounded gravels to cobbles in a matrix of sand, silt, and clay. Found and mapped in the west side of the South Mountain Quadrangle at base of the Stansbury Range and at the mouth of Welch Canyon, Cow Hollow, and other minor drainages. Deposits tend to be much coarser (cobble) in the headland portion of the fans and become thinner and finer grained (coarse sand – pebble) in the more distal portion of the fan slope. The fans overlie and incise older fans (Qaf_3)

but are also incised by lacustrine processes related to the occupation of Lake Bonneville (both the transgressive and regressive portions of the lake cycle) in the basin; therefore, deposition of the fans is either coeval with or predate the lake. Maximum thickness of the deposits is expected to be less than 40 feet (12 m).

Qaf_3 – level 3 alluvial fan deposits, pre Bonneville lake cycle (upper to middle Pleistocene). Very poorly to moderately sorted, subangular to rounded gravels to cobbles in a matrix of sand, silt, and clay. Found and mapped in the west side of the South Mountain Quadrangle at base of the Stansbury Range and in the southwest corner of the Stockton Quadrangle. Fan deposits in the Stockton Quadrangle originate from a large fan from Ophir Canyon to the south, whereas the deposits in the South Mountain Quadrangle were emitted from the mouths of local drainages of the Stansbury Mountains. The large alluvial fans on the Stansbury range near East Hickman Canyon have also been potentially classified as pediment surfaces by Rigby (1958). Bedrock outcrops are seen in multiple drainages in the area and may indicate that the alluvial cover is only a thin veneer and that some of these features would be better classified as older piedmont surfaces. Qaf_3 is deeply incised in multiple localities by ephemeral streams (Qal) and normal faults from basin and range extension. Maximum thickness of the deposits is believed to be less than 40 feet (12 m).

Qaf_o – older alluvial fan deposits, undivided (upper Pleistocene to upper Miocene). Pre-Bonneville fans that are poorly to moderately sorted, subangular to rounded gravels to cobbles in a matrix of sand, silt, and clay. Qaf_o deposits may be composed of Qaf_3 and QTaf older deposits; however, they are undivided because the specific contacts of the deposits could not be determined. The deposits tend to be much coarser (cobble) in the headland portion of the fans and become thinner and finer grained (coarse sand – pebble) in the more distal portion of the fan slope. The fans are extensively incised by modern drainage patterns but these modern deposits are not significant enough to be mapped at this scale. The fans have also been reworked by Bonneville age lacustrine processes in the more distal portions of these deposits. Maximum thickness of the deposits is expected to be less than 100 feet (30 m).

QTaf – Oldest alluvial fan deposits (Middle Pleistocene to upper Miocene). Pre Bonneville fans that are unconsolidated to partially consolidated, poorly to moderately sorted sub angular to rounded gravels to cobbles in a matrix of sand, silt, and clay originating from debris flows. The fans are deeply incised by modern drainage systems; however, most of these modern deposits were not considered substantial enough to be mapped at this scale. Tooker and Roberts (1992) mapped this unit as Harkers alluvium. The unit is described as a fanglomerate, comparable to very old alluvial fan structures on the eastern side of the Oquirrh Range near Magna, Utah (Tooker and Roberts, 1971; Slentz, 1955). Tooker and Roberts (1992) considered the deposit as being early Pleistocene; however, Solomon (1993) and Black and others (1999) renamed the fan as QTaf, and classified it to be as early as late Tertiary, due to its stratigraphic position between Tertiary volcanics of the Salt Lake Formation and the more modern alluvial fans in the area. Topographic relief between the deep incision in Soldiers Canyon and the top of the fan is over 200 ft; however, the actual thickness of the deposits is unclear. Estimated maximum thickness of the deposits are ~200 feet (~60 m).

Lacustrine Deposits

Qlgb – Lacustrine gravel and sand, related to the Bonneville level of Lake Bonneville (upper Pleistocene). Moderately to well-sorted clast supported gravels and occasional cobbles within a matrix of coarse to medium sand and occasional fine sand. Gravel clasts are subangular to rounded due to the proximity of bedrock sources or the reworking of local alluvial fans. Gravel deposits in the Stockton quadrangle are reworked from QTaf deposits of the Soldier Canyon Fan and are better developed near the mouth of Soldier Canyon. Gravel deposits become thinner towards the south along this shore face. The only other location where prominent gravel deposits accumulated at the Bonneville level is near the mouth of Welch Canyon. Maximum thickness of the deposits is less than 40 feet (12 m).

Qlg – lacustrine gravel and sand, undivided (upper Pleistocene). Moderately to well sorted clast supported gravels and occasional cobbles within a matrix of coarse to medium sand and locally derived fine sand. Gravel clasts are subangular to subrounded due to the proximity of bedrock sources or reworking of local alluvial fans. Thickness of the deposits varies ranging from thin 3 feet (1 m) to ~40 feet (12 m).

Qlf – lacustrine mud (Holocene to upper Pleistocene). Silt, muds, and clay with subordinate amounts of sand, deposited in relatively quiet waters of Rush Lake and Lake Bonneville. Thickness of the deposits are estimated to be less than 20 feet (6 m).

Marsh and Spring Deposits

Qsm – marsh deposits (Holocene). Organic rich clay, silt, and fine grained sand related to a shallow groundwater table, spring outflow and channels. Springs flow in low lying areas around the valley and are concentrated near the modern day Rush Lake, within fine grained lacustrine muds (Qlg) and undifferentiated lacustrine and alluvial deposits (Qla). Maximum thickness of the deposits is less than 10 feet (3 m).

Mass Movement Deposits

Qms – landslide deposits (Holocene to upper Pleistocene). Irregular, unconformable slide blocks of detached and rotated Mississippian Great Blue Limestone and Manning Canyon Shale that are present on the south side of Soldier Canyon, southeast of Stockton (Tooker and Roberts, 1992). The upper scarp is within the map area, whereas the majority of the slide blocks are located just north of the map area.

Mixed Environment Deposits

Qac – alluvium and colluvium, undivided (Holocene and latest Pleistocene). Unconsolidated alluvium and colluvium on gentle slopes composed of primarily sandy deposits and lesser amounts of boulders, gravel, silt and clay. Maximum thickness of the deposits is expected to be less than 20 feet (6 m).

Qla – lacustrine and alluvial deposits, undivided (Modern to upper Pleistocene). Undifferentiated deposits of calcareous marl, silt, sand, and gravel consisting of lacustrine and alluvial origin. These deposits include pre Bonneville alluvial deposits reworked or etched by lacustrine processes during the occupation of Lake

Bonneville or lacustrine deposits that cannot readily be determined at the map scale. Deposits vary in thickness and are estimated to be no thicker than 6 to 12 feet (2 to 4 m).

Stacked Unit Deposits

Qla/Qaf$_2$ – undivided lacustrine and level 2 alluvial fan deposits of Bonneville age (Holocene to upper Pleistocene). A thin veneer of undifferentiated lacustrine and alluvial deposits overlying alluvial fan deposits, deposited prior to and during the occupation of the lake in the basin. The unit is mostly likely due to the reworking of Qaf$_2$ deposits by lacustrine processes associated with the transgressive rise of Lake Bonneville, and the reworking of these deposits by alluvial processes after the regression of Lake Bonneville. Qla deposits are very thin (< 3-6 feet (1-2 m)) and patchy due to erosion by alluvial sheetwash and other overland flow that caused rills and gullies.

Qla/Qaf$_3$ – undivided lacustrine and alluvial deposits over older alluvial fan deposits (Holocene to middle Pleistocene). A thin veneer of undifferentiated lacustrine and alluvial deposits overlying older alluvial fan deposits. Lacustrine processes associated with the transgressive rise of Lake Bonneville reworked older Qafo deposits into small patches of lacustrine sands, gravels, and silts. Following the regression of Lake Bonneville, additional alluvial processes have eroded and reworked the sediments causing small rills and gullies in the unit. Qla deposits are very thin (< 3-6 feet (1-2 m) and patchy due to erosion by sheetwash and other overland flow.

Qla/QTaf – undivided lacustrine and alluvial deposits over older alluvial fan deposits (Holocene to upper Miocene). The unit consists of a thin veneer of undifferentiated lacustrine and alluvial deposits overlying the oldest alluvial fan deposits.

Lacustrine processes associated with the transgressive rise of Lake Bonneville reworked older QTaf deposits and deposited small patches of lacustrine sands, gravels, and silts. Following the regression of Lake Bonneville, additional alluvial processes have eroded and reworked the sediments causing small rills and gullies in the unit. Qla deposits are very thin (< 3-6 feet (1-2 m) and patchy due to erosion by overland flow.

Qlf/Qalo – lacustrine mud (related to Lake Bonneville) over older Rush Valley alluvium (Holocene to mid Pleistocene). The unit consists of a thin veneer of fine grained lacustrine muds, silts, and calcareous marl deposited in relatively deep and quiet waters of Lake Bonneville. Qalo deposits are thin (< 6 feet (2 m)). Qlf deposits overlie older alluvium and are composed of moderately sorted interbedded gravels to cobbles within a matrix of sands, and silts deposited in channels, floodplains, and minor terraces that developed prior to the occupation of Lake Bonneville in the basin. Thickness of the Qalo deposits is unknown.

Bedrock Units

Bx – pre-Quaternary bedrock – (Pliocene to Cambrian). Bedrock in the quadrangle has been mapped as undifferentiated bedrock. The valley is bounded on the east and west by Permian, Pennsylvanian, and Mississippian carbonates and quartzites of the Oquirrh Group and Great Blue Limestone formation. South Mountain dominates the center of these two quadrangles and is composed of similar bedrock units as the Stansbury and Oquirrh ranges; however, the mountain also hosts some small intrusions of Tertiary volcanics (Gilluly, 1932). The interested reader can see Gilluly (1932), Moore and Sorensen (1979), Tooker (1980), and Tooker and Roberts (1988a, 1988b, 1992) regarding the bedrock within South Mountain and the Oquirrh Mountains to the east and Moore and Sorensen (1979) or Rigby (1958) regarding the Stansbury Mountains to the west.

Acknowledgements

The authors thank Dr. Jack Oviatt for his assistance and thoughtful discussion in the field and Don Clark for his thorough reviews of the map and manuscript. This fieldwork was supported by the United States Geological Survey's EDMAP program (EDMAP Agreement No. 07HQAG0046) and the National Science Foundation (NSF 1053129) funded the radiocarbon dates.

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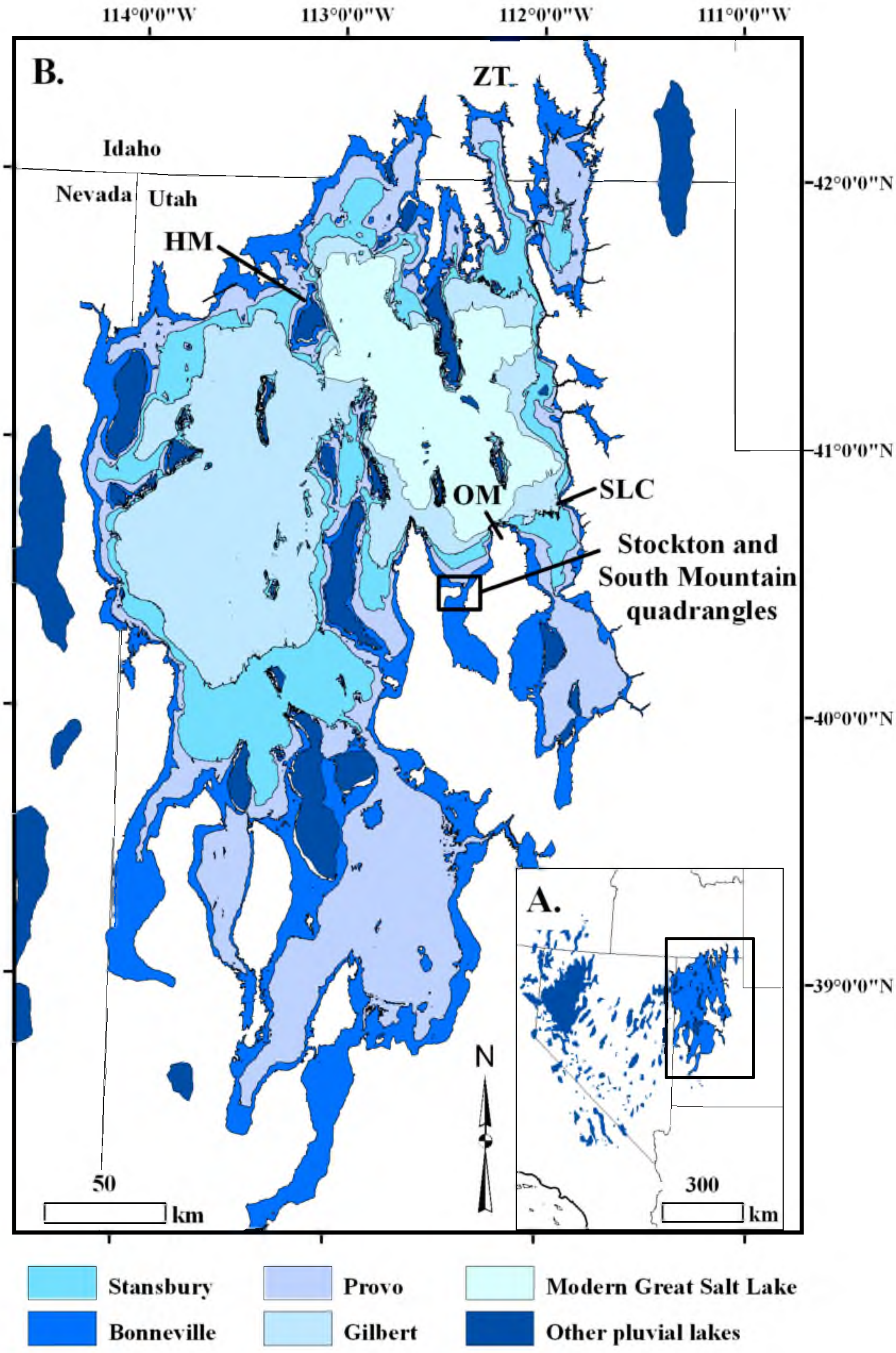
Figure 2.1: Lake Bonneville maximum extent of major lake levels. A) Regional map of the extent of Lake Bonneville at its maximum in relation to the maximum of other pluvial lakes in the Great Basin; B) The maximum extent of the modern Great Salt Lake and the maximum extent of each of the major lake levels of the Lake Bonneville lake cycle. Extent of Lake Bonneville at the Bonneville level in relation with the South Mountain and Stockton quadrangles.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B.
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South Mountain
quadrangles
42°0'0"N
41°0'0"N
40°0'0"N
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Provo
Modern Great Salt Lake
Bonneville
Gilbert
Other pluvial lakes

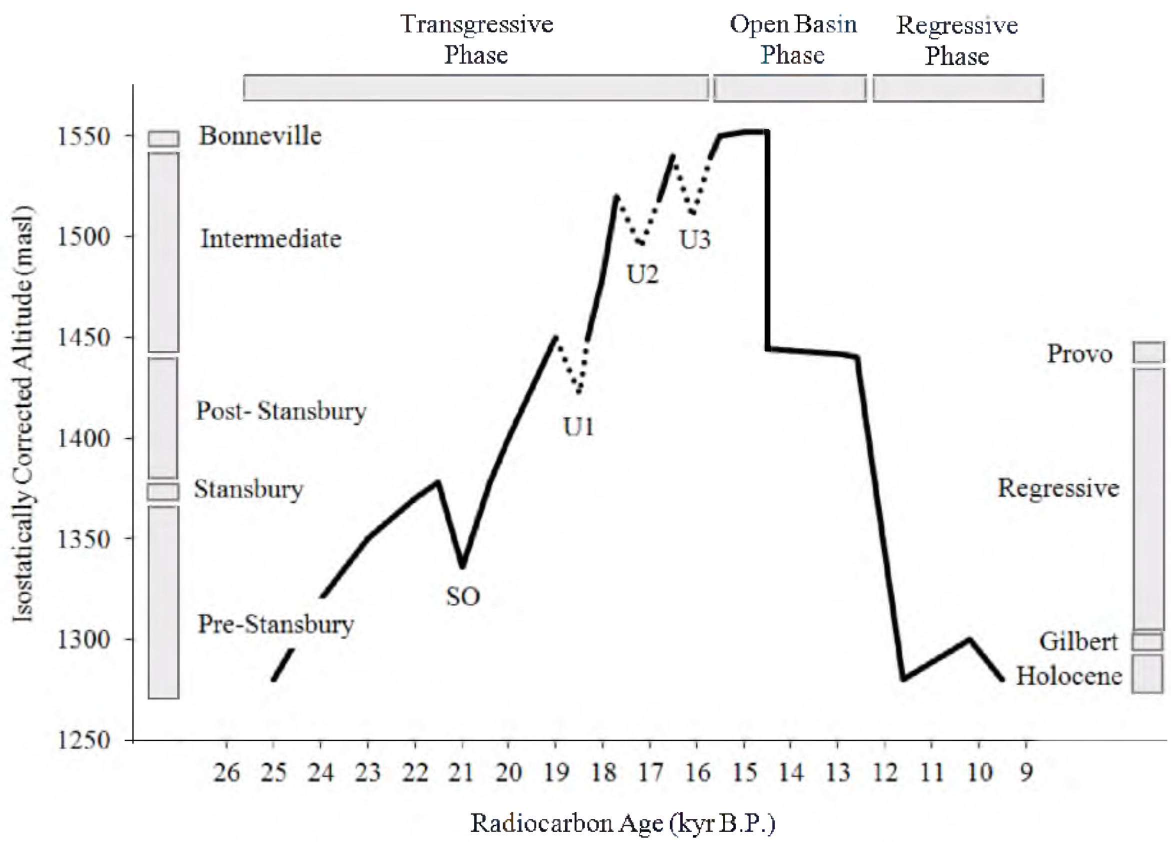


Figure 2.2: Lake Bonneville hydrograph modified from Oviatt (1997) and Godsey et al. (2011). Altitudes are adjusted for effects of differential isostatic rebound in the basin (Oviatt et al., 1992). Amplitude limits of lake stage fluctuations associated with the U1, U2, and U3 oscillations are approximate and are shown here schematically. The temporal range of the transgressive, regressive, and open phases of the lake cycle are shown horizontally, whereas the altitudinal range of each of the paleoshorelines is shown vertically within either the transgressive or the regressive phases of the lake cycle.

Figure 2.3: 2 m auto correlated DEM hill shade with mapped paleoshoreline features. Paleoshorelines (1-4) are Intermediate paleoshoreline features north of the Stockton threshold; paleoshorelines (a-j) are Intermediate paleoshorelines within the Stockton threshold; paleoshorelines (B1-B4) are paleoshoreline features associated with the Bonneville level; and paleoshorelines (Z, Y, and Rh) are the Shambip, Smelter, and Holocene Rush Valley Paleoshorelines.

Figure 2.4: Field photographs of sampled radiocarbon localities. White bar for scale and stars for sample locations. See Plate 2.1 for sample locations. Photo A) Shambip gravel bar at Indian Hill (T. 5 S., R. 5. W., Sec. 8) with fine grained lacustrine muds and sands deposited behind the bar. The gravels have been locally mined. White box is extent of photograph B. Photograph is being taken towards the Oquirrh Mountains to the west and Indian Hill at my back. Photo B) Blowup of radiocarbon sample site at Indian Hill. Sample Beta-307252 (*Stagnicola bonnevillensis*) and Beta-307253 (*Valvata utahensis*) was sampled from a gravely sand and obtained ages of 13,300 & 13,360 ^{14}C B.P., respectively. Photo C) Shambip offshore gravels and sands on south side of South Mountain (T. 4 S., R. 5. W., Sec. 33). Sample location within septic system pit on southern the south side of South Mountain. Freshwater gastropod shells, Beta-307254 (*Valvata utahensis*), were collected and analyzed from fine offshore sands to obtain constrain the Shambip paleoshoreline at an age of 14,290 ^{14}C B.P.

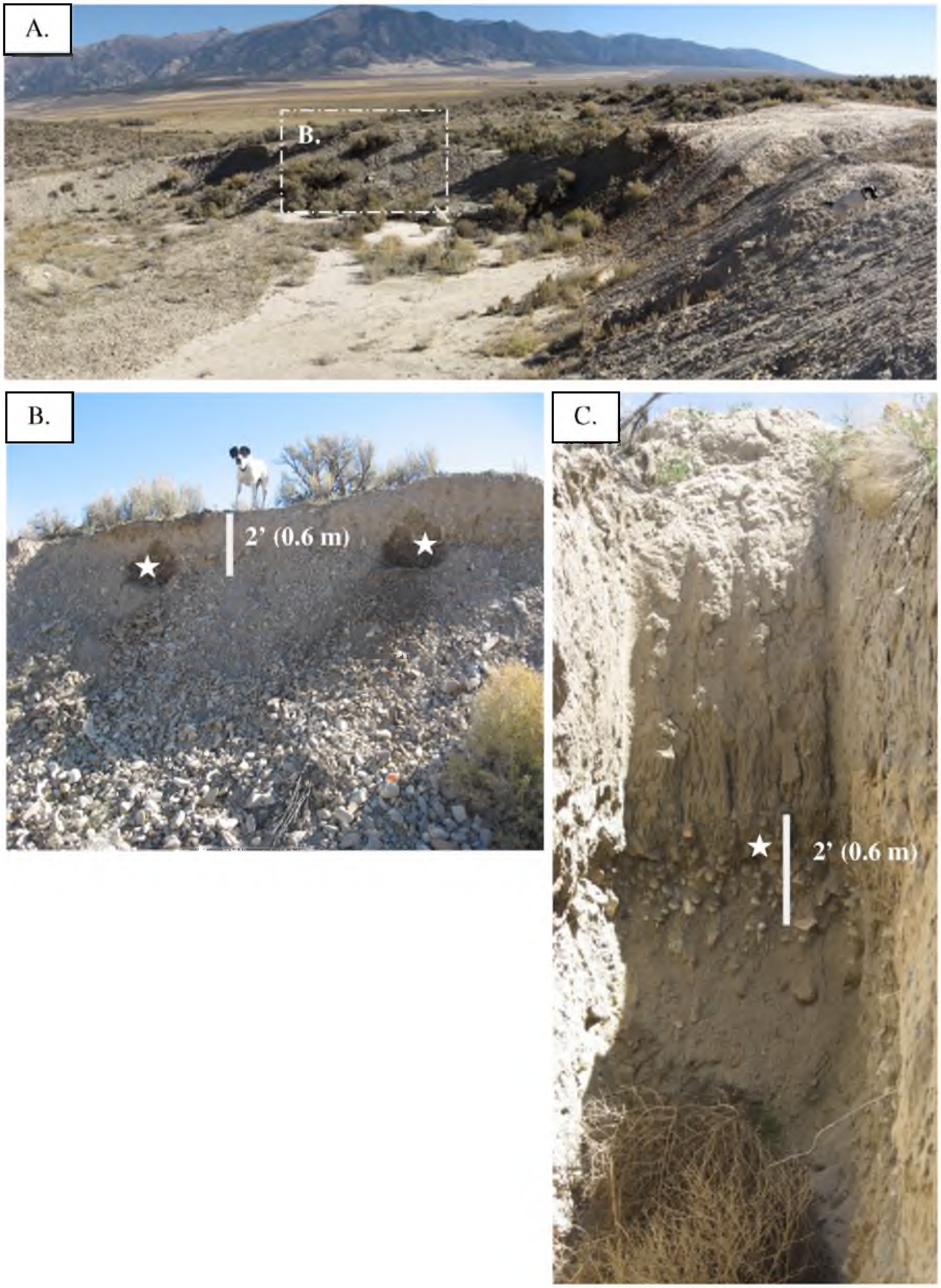

A.
B.
B.
2’ (0.6 m)
C.
2’ (0.6 m)

Table 2.1: Lake Bonneville and Rush Valley paleoshoreline ages.

Ages of major paleoshorelines of Lake Bonneville and Rush Valley and shoreline elevations of these shorelines within the map area.

Lake Cycle and Phase	Paleoshoreline (map symbol)	Radiocarbon Yrs BP	Local altitude - feet (meters)	Isostatically corrected altitude - feet (meters)
Lake Bonneville				
Transgressive	Stansbury	22,000 - 20,000[1]	Not Present	Not Present
	Bonneville (B)	15,500 - 14,500[2]	5233-5250 (1595-1600)	5077-5092 (1547-1552)
Regressive	Shambip (Sh)	14,500 - 13,300 (?)[3]	5052-5068 (1540-1545)	4909-4924 (1496-1500)
	Provo (main basin)	14,500 - 12,000[4]	Not Present	
	Smelter (Sm)	12,000 - 10,000 (?)[5]	5003-5020 (1525-1530)	4863-4878 (1482-1487)
	Gilbert (main basin)	11,000 - 10,000[6]	Not Present	
Holocene	Rush Lake (R)	10,000 (?) - Present	4954-4977 (1510-1517)	4817-4838 (1468-1475)

[1]Oviatt and others. (1990). Currey (written communication to the Utah Geological Survey, 1996) assumed a maximum age for the Stansbury paleoshoreline of 21,000 ^{14}C yr B.P.

[2]Oviatt and others. (1992), Oviatt (1997)

[3]Estimated duration of the paleoshoreline may be extended to 12,000 ^{14}C yr B.P (Godsey et al., 2011); however, current dates are constrained to reported values.

[4]Godsey and others (2005) revised the timing of the occupation of the Provo shoreline and subsequent regression; Oviatt and others (1992) and Oviatt (1997) proposed a range from 14,500 to 14,000 ^{14}C yr B.P. Oviatt and Thompson (2002) summarized many recent changes in the interpretation of the Lake Bonneville radiocarbon chronology.

[5]Estimated date based on assumption that the paleoshoreline is equivalent to the Gilbert age lake within the main body of Lake Bonneville (Oviatt and others., 2005); estimated the age of the Rush Valley to be less than 10,000 ^{14}C yr B.P.

[5]Calendar calibrated age of the end of the Provo shoreline estimated by interpolation from data in Godsey and others (2005), table 1, who used Stuiver and Reimer (1993) for calibration.

[6]Murchison (1989), figure 20.

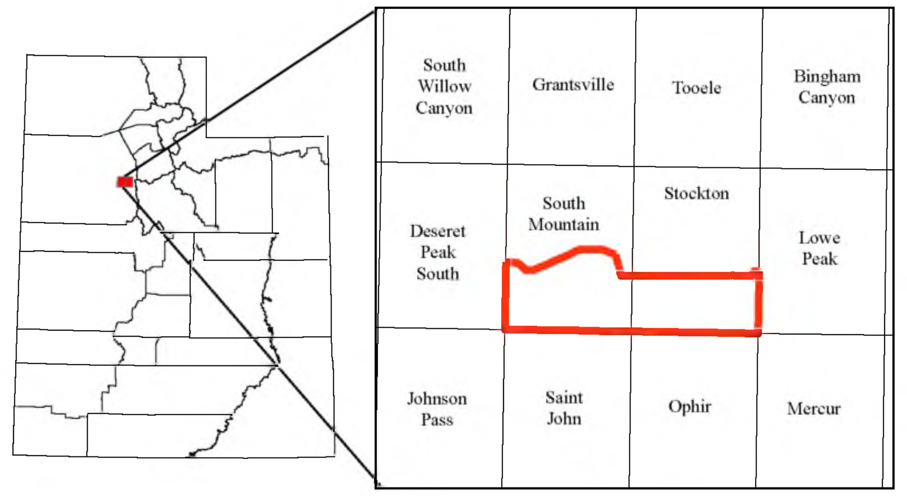
Plate 2.1: Surficial geologic map of southern portions of the Stockton and South Mountain quadrangles, Tooele County, Utah

Plate 2.1

GEOLOGIC MAP OF UNCONSOLIDATED DEPOSITS IN THE SOUTHERN PORTIONS OF THE SOUTH MOUNTAIN AND STOCKTON QUADRANGLES, TOOELE COUNTY, UTAH

by

Daren T. Nelson and Paul W. Jewell



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1:24,000

0 0.25 0.5 1 1.5 2 Kilometers

0 0.25 0.5 1 1.5 2 Miles

Base map from U.S. Geological Survey
Stockton and South Mountain 7.5' quadrangles, 1993
Geologic data and base map in NAD 1927

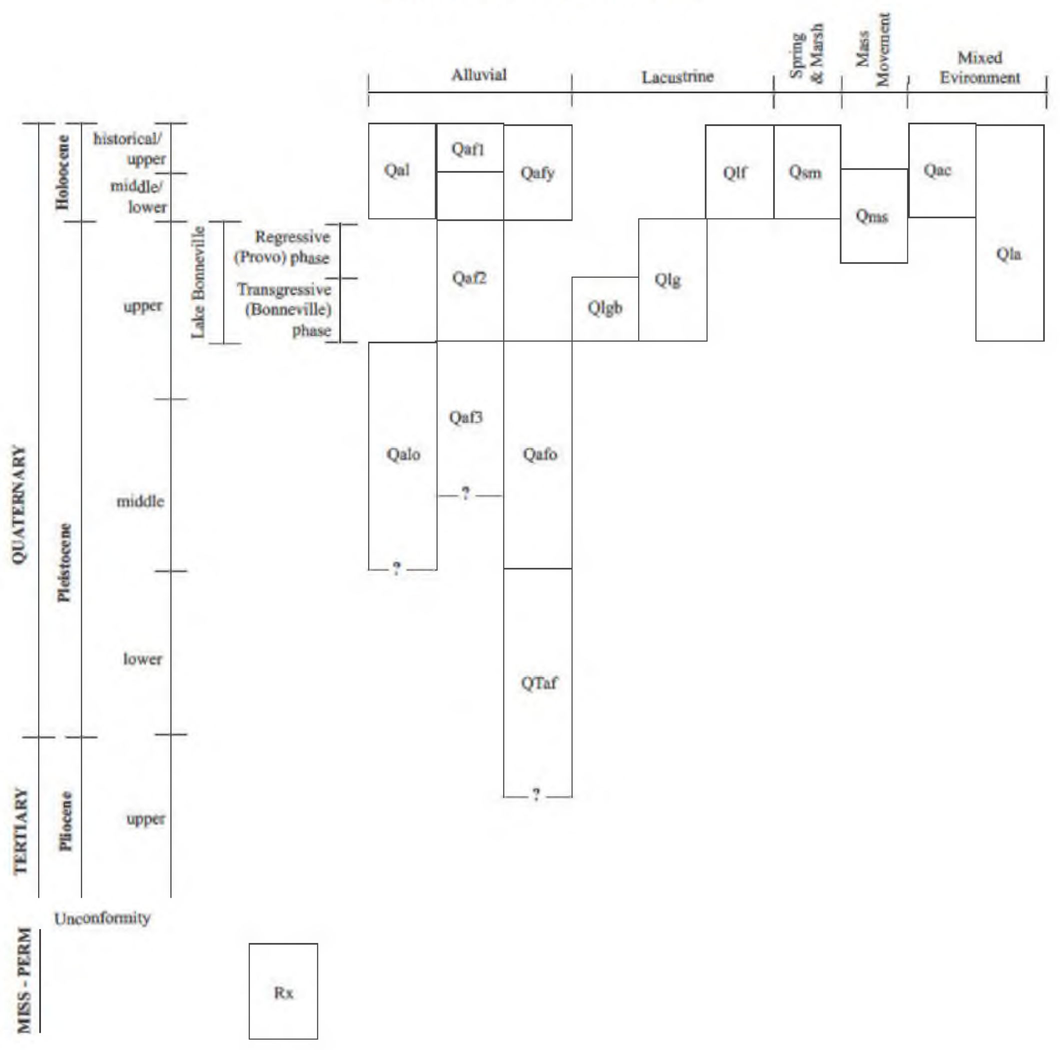
The geologic map was funded by the U.S. Geological Survey (EDMAP program)
and a Graduate Research Grant from the Geological Society of America.

This map is a draft and is under review with the Utah Geological Survey.

- Rx
- Ms
- Qaf1
- Qaf2
- Qaf3
- Qafy
- Qafo
- QTaf
- Qal
- Qlf
- Qlg
- Qlgb
- Qsm
- Qac
- Qla
- Qla/Qaf2
- Qla/Qaf2
- Qla/Qafo
- Qla/QTaf
- Qlf/Qalo

- Holocene-Rush Paleoshoreline
- Bonneville Paleoshoreline
- Bonneville Paleoshoreline (~ location)
- Bonneville Paleoshoreline (inferred)
- Shambip Paleoshoreline
- Shambip Paleoshoreline (~ location)
- Smelter Barrier Ridge
- Smelter Barrier Ridge (~ location)
- Smelter Paleoshoreline
- Smelter Paleoshoreline (~ location)
- Contact
- Contact (~ location)
- Landlide Scarp
- Normal Fault (~ location)
- Normal Fault
- Lineament, Proposed Fault
- Map Boundary,
- Solomon (1993) lines

CORRELATION OF MAP UNITS



CHAPTER 3

GEOLOGIC MAP OF UNCONSOLIDATED DEPOSITS IN THE HOGUP BAR QUADRANGLE, BOX ELDER COUNTY, UTAH

Location and Geographic Setting

The Hogup Bar quadrangle is in the northwestern portion of the Bonneville basin, a subbasin of the Basin and Range Province (Plate 3.1). The quadrangle is near the northwestern shores of the Great Salt Lake, a remnant of late Pleistocene Lake Bonneville, ~ 15 miles (24 km) southeast of Park Valley, Utah and ~ 87 miles (140 km) northwest of Salt Lake City, Utah.

Historically, the transcontinental railroad crossed the northwestern corner of the quadrangle where the small loading station of Ombey was located. This railway was active from 1869 to 1942; however, the construction of the Lucin Cutoff in 1904 shortened the length of the railroad by crossing the Great Salt Lake and made this section of the railway practically obsolete (Huchel, 1999). The railway was in working order until 1942 when it was dismantled to provide steel for the war effort (World War II) (Huchel, 1999). The railway grade is now a backcountry Scenic Byway maintained by the U.S. Bureau of Land Management, and the primary use for the quadrangle is winter range for sheep and cattle.

Hydrologic features in the quadrangle include multiple ephemeral gullies that drain either into internal closed basins, such as the West Desert or Dove Creek sub basins, or directly into the Great Salt Lake. The two largest ephemeral streams within the quadrangle are Dove Creek, located in the northwestern corner of the quadrangle, and an unnamed drainage emanating from Big Pass in the southeastern corner of the map.

Previous Geologic Studies

Stifel (1964) mapped the bedrock and surficial geology of Terrace and Hogup Mountains, including the Hogup Bar quadrangle, at a scale of 1:63,360. Stifel's (1964) analysis of the area focused on the bedrock geology and did not provide a detailed analysis of the Quaternary age deposits. Stifel mapped the Quaternary deposits based on clast size, and did not distinguish specific landforms. Doelling (1980) subsequently mapped the geology of Box Elder County at a scale of 1:125,000. Ongoing geologic mapping by the United States Geological Survey (USGS) and the Utah Geological Survey (UGS) includes maps at a 1:100,000 scale by Miller and Felger, in review and Miller et al., in prep. In addition to these broader geologic maps, Cavas (2003) mapped the adjacent Matlin quadrangle and McCarthy and Miller (2002) and Miller and McCarthy (2002) mapped the Terrace Mountain East and West quadrangles west of the study area at a scale of 1:24,000.

Previous investigators recognized the importance of the Quaternary geomorphic features within the area. G.K. Gilbert (1890) discussed the Dove Creek area in his investigation of what he termed the "Intermediate shorelines" (paleoshorelines formed during the transgression of Lake Bonneville between the altitudinal range of the

Bonneville and Provo paleoshorelines) that straddle the Matlin and Hogup Bar quadrangles. For this publication, Gilbert's "Intermediate shorelines" are termed the Intermediate paleoshorelines.

Gilbert (1890) discussed how the crests of paleoshorelines of the Bonneville age lake exhibit multiple altitudes in a variety of locations within the basin, and he identified the Hogup - Matlin area as an example of this phenomenon. Godsey et al. (2005) also discussed multiple levels of the Provo age lake in the Hogup Bar quadrangle. Multiple radiocarbon dates and stratigraphic evidence suggest that the lake level stabilized near the Provo until ~ 12,600 ^{14}C yr. B.P. (Godsey et al., 2011). Cavas (2003) and Sack (1999) also discussed the Hogup Bar and Matlin quadrangles as localities exhibiting transgressive deposits just below the Provo level, which they call the sub Provo shorezone. Within this study, features that Cavas (2003) and Sack (1999) term the sub Provo paleoshorelines are included in a subset of paleoshoreline deposits termed the Post Stansbury paleoshorelines.

General Quaternary Geology

The excellent Pleistocene record within the Bonneville basin indicates that the Great Salt Lake is a remnant of a series of large pluvial lake systems that have occupied the basin since the middle Pleistocene (Scott et al., 1983; McCoy, 1987). Lake Bonneville was the most extensive and recent of these lake cycles. The initial rise of this lake occurred ~ 27,000 ^{14}C yr B.P., reached its maximum extent ~ 15,300 ^{14}C yr B.P., and lowered to its modern lake level ~ 10,000 ^{14}C yr B.P. (Oviatt et al., 1992; Oviatt, 1997). Lake Bonneville had four major periods when significant shorelines developed, known today from oldest to youngest as the Stansbury, Bonneville, Provo, and Gilbert paleoshorelines (Fig. 3.1). The lake at its maximum extent, the Bonneville level, encompassed ~51,556 km^2 (updated calculation based on the methodology by Wambeam (2001) and Nelson and Jewell (in review, Chap. 4) of northwestern Utah, portions of southern Idaho, and eastern Nevada and a volume of ~10,494 km^3.

The hydrologic budget and resultant lake level of the terminal basin were susceptible to continuous climatic and seasonal fluctuations during the closed period of the lake's transgressive phase (~27,000 – 15,300 ^{14}C yr B.P.) and regressive phase (~12,600 – 10,000 ^{14}C yr B.P.) (Fig. 3.1). Besides small seasonal lake level fluctuations there are multiple larger (20-45 m) oscillations during the transgression of the lake. The Stansbury oscillation (~ 20,000 – 21,000 ^{14}C yr B.P.) is the most prominent and well studied oscillation (Oviatt et al., 1990); however, up to five (5) additional large oscillations have been proposed during the transgression of the lake (Currey and Oviatt, 1985; Oviatt, 1997; Nelson and Jewell, in review, Chap. 4).

The water level continued to rise in the basin until the lake started to overflow a topographic divide near Zenda, Idaho ~ 15,300 ^{14}C yr B.P. (Oviatt et al., 1992; Oviatt, 1997). The Zenda threshold is in a narrow valley with thick deposits of unconsolidated fan materials overlying the semiconsolidated strata of the Tertiary Salt Lake Formation (Janecke and Oaks, 2011; Thackray et al., 2011). The lake catastrophically breached this topographical divide and flooded into the Columbia River drainage at ~ 14,500 ^{14}C yr B.P. (O'Conner, 1993). This flood caused a regression that dropped the lake by ~ 350 feet (~108 m), until the lake elevation was constrained by the continued overflow of a bedrock controlled topographical divide near Red Rock Pass, Idaho (Janecke and Oaks, 2011; O'Conner, 1993). The bedrock threshold kept the water level of the lake relatively stable for ~ 2,000 – 2,500 yrs, to form the paleoshorelines of the Provo level, until the lake continued to regress beginning ~ 12,600 ^{14}C yr B.P. (Godsey et al., 2011). The regression of the water level following the Provo age was very rapid and may be related to the Bøelling Allreød warming event (Benson et al., 2011; Godsey et al., 2011). Following this rapid regression, a small oscillation coinciding with the timing of the Younger Dryas climatic event has been correlated with paleoshoreline development of the Gilbert level (Oviatt et al., 2005).

Late Pleistocene Paleoshoreline Features

Within the Hogup Bar quadrangle, there are numerous depositional and erosional paleoshorelines. All four major paleoshorelines (Stansbury, Bonneville, Provo, and Gilbert) are exhibited in the quadrangle; however, there are also more than 35 lesser but distinct paleoshorelines. Paleoshorelines in the Hogup Bar are classified into eight distinct groups based on their relative position/altitude, relative age, and stratigraphic relationships (Table 3.1). The four major paleoshoreline features and the largest Intermediate paleoshoreline features have been mapped (Plate 3.1). Smaller paleoshorelines exist in the map area, but are not distinctly mapped.

Gilbert (1890) was the first to refer to a set of paleoshorelines located between the more prominent Provo and Bonneville paleoshorelines as the "Intermediate shorelines." However, there is evidence of many other paleoshorelines that are lower in altitude than these "Intermediate "paleoshorelines. These lower paleoshorelines are located between the Provo and Stansbury shore zones. Therefore, it is necessary to distinguish the lower transgressive paleoshorelines located between the Provo and Stansbury shore zones from the upper transgressive paleoshorelines located between the Provo and Bonneville shore zones. The authors will follow the terminology of Gilbert by referring to the upper paleoshorelines as the Intermediate paleoshorelines; and the lower intermediate paleoshorelines as Post Stansbury paleoshorelines. These features will be called the Post Stansbury paleoshorelines since they were deposited during the transgressive phase of the lake following the formation of Stansbury age paleoshorelines.

Pre-Stansbury Paleoshorelines

Transgressive paleoshorelines exhibited by multiple gravel (Qlg) berms and beaches partially buried by lacustrine marl (Qlm). Dove Creek incised prominent deposits associated with these paleoshorelines (sec. 32, T. 10 N., R. 12 W., S. 32, Salt Lake Base Line and Meridian [SLBM]) and smaller ephemeral streams expose these paleoshorelines near Peplin Flats (sec(s) 1 and 2, T. 10 N., R. 12 W., SLBM).

Stansbury Paleoshorelines

Transgressive paleoshorelines exhibited as wave platforms or as multiple gravel (Qlg) barrier ridges or berms partially buried by lacustrine marls (Qlm). Stansbury landforms are best represented as a zone of paleoshorelines rather than a single paleoshoreline at a distinct altitude. Transgressive features within this zone consist of both erosional and depositional landforms deposited during a large oscillation (150 feet [45 m]) of the lake’s water level (Oviatt, 1987; Oviatt et al., 1990; Currey et al., 1985; Patrickson et al., 2011). Within the Hogup Bar quadrangle, landforms in the Stansbury zone are very similar to other localities in the basin (Oviatt, 1991). These deposits usually consist of gravel barriers and erosional and/or depositional wave platforms with carbonate encrusted sand deposits, tufa encrusted gravels (beachrock), occasional tufa mounds, and occasional charophyte debris partially buried by fine grained lacustrine marls (Qlm) deposited during the transgression of the lake (Fig. 3.2).

Post-Stansbury Paleoshorelines

Transgressive gravel and sand (Qlg) deposits comprised of multiple paleoshorelines between the altitude of the Stansbury shorezone and the Provo shorezone. Many paleoshorelines are partially buried by Qlm deposits and regressive paleoshorelines associated with the Provo level of the lake. Sack (1999) and Cavas (2003) suggest that significant paleoshorelines formed during the transgression of the lake near the altitude of the Provo level. The most significant post Stansbury paleoshoreline in the quadrangle is at an altitude of ~ 4,646 feet asl (~1,416 masl). The paleoshorelines are exhibited as multiple substantial baymouth bars near Dove Creek (sec(s) 9, 10, and 17, T. 10 N., R. 12 W., SLBM) and in the northern portion of the quadrangle near Peplin Flats (sec(s) 10 and 11, T. 10 N., R. 12 W., SLBM).

Intermediate Paleoshorelines

Transgressive gravel and sand (Qlg) deposits make up multiple erosional and depositional paleoshorelines between the altitude of the Provo and the Bonneville shore zones. As the lake rose, transgressive marls (Qlm) partially buried the transgressive gravel and sand deposits (Qlg) deposits; but most of these finer grained deposits have been eroded away following the regression caused by the Bonneville flood. At least six prominent and four less distinct Intermediate paleoshorelines can be found in the quadrangle. Many of these paleoshorelines exhibit a range of altitudinal limits and are not always preserved from location to location. Gastropod samples from a transgressive sand below the Provo platform in Big Pass (Sample 1: Table 3.2) suggest that the Intermediate paleoshorelines started to form ~19,000 ^{14}C B.P. (Sack, 1999). Within the study area, additional evidence for at least three oscillatory events during the development of the Intermediate paleoshorelines exists on the western flanks of Hogup Mountain (sec. 18, T. 10 N., R. 11 W., SLBM), two of which were previously unknown (Nelson and Jewell, in review, Chap. 5). Gastropod samples collected from deposits related to these oscillatory events have reported ages from 16,500 – 18,900 ^{14}C B.P. (Samples 2-9: Table 3.2). These paleoshorelines are difficult to trace in other localities of the Bonneville basin (Gilbert, 1890; Nelson and Jewell; in press). The most prominent localities where these paleoshorelines are preserved are within an embayment west of Big Pass (sec(s) 1, 2, and 11-14, T. 9 N., R. 12 W., SLBM – Fig. 3.3) and southeast of Hogup Mountain (sec(s) 24 and 25, T. 10 N., R. 12 W., and sec(s) 19 and 30, T. 10 N., R. 11 W., SLBM – Fig. 3.4).

Bonneville Paleoshorelines

Transgressive gravel and sand (Qlg) deposits make up multiple depositional and erosional paleoshorelines. Bonneville age paleoshorelines are the most recognizable paleoshoreline due to their distinct contrast with non lacustrine deposits. Bonneville age paleoshorelines are also more prominent than the other transgressive paleoshorelines because the water level stabilized, for ~ 1,000 yrs as the lake overflowed into the Snake River/Columbia River basin near Zenda, Idaho (Table 3.1) (Oviatt, 1997). Bonneville age lacustrine landforms include spits, baymouth bar complexes, wave platforms, and tombolos. The most prominent Bonneville age feature within the quadrangle is a substantial spit on the south side of Hogup Mountain referred to by the authors as the Hogup Spit (sec. 19, T. 10 N., R. 11 W., SLBM) (Fig.3.4).

Provo Paleoshorelines

Regressive gravel and sand (Qlg) deposits overlying marl (Qlm) deposits and comprises multiple depositional paleoshorelines and erosional platforms (Table 3.1). Provo age lacustrine sediments tend to be finer grained than Bonneville age lacustrine sediments. Provo age sediments were derived from reworked lacustrine source material, whereas Bonneville age sediments were usually derived from reworked colluvial and/or alluvial deposits. Provo age paleoshorelines are the second most recognizable paleoshoreline in the area and are usually more extensive (laterally and vertically) than Bonneville age paleoshorelines. Godsey et al. (2005) suggest that at least nine distinct paleoshorelines associated with the Provo level, and at least five of these are present within the Hogup Bar Quadrangle. Two Gastropod samples (Samples 10 and 11: Table 3.2) collected from the lowest Provo age deposits, locally have ages of 11,910 and 12,430 ^{14}C B.P (Godsey et al., 2005). Geomorphic features associated with the Provo level include multiple spits, beach barriers, baymouth bars, wave platforms, and tombolos (Fig. 3.5). The most prominent Provo features in the quadrangle are the substantial wave platforms excavated on the northern and northeastern slopes of Hogup Mountain (sec. 19, T. 10 N., R. 11 W., SLBM). Tufa caps are prevalent on the basinward side of the platforms, whereas erosional notches are prevalent near the cliff faces.

Regressive Paleoshorelines

Regressive gravel and sands (Qlg) deposits overlie fine grained lacustrine marl and silt (Qlm), transgressive gravel and sand (Qlg), and transgressive sand (Qls) deposits. These paleoshorelines are not as common as transgressive paleoshorelines, suggesting that the water level lowered quickly and dramatically following the occupation of the lake at the Provo level. The quick regression of the lake may correlate to the warming period of the Bølling-Allerød interstadial (Benson et al., 2011; Godsey et al., 2011).

Gilbert Paleoshorelines

Gravel and sand (Qlg) deposits making up two small beach berms in the northeastern portion of the quadrangle (sec(s). 5, 6, and 8, T. 10 N., R. 11 W., SLBM). The small berms are near the altitudinal extent of other Gilbert age deposits in the basin and do not have overlying marl deposits. It is suggested that these features are landforms related to the Gilbert age lake; however, there is no radiometric dating to support this interpretation. The Gilbert oscillatory event may be connected with the Younger Dryas event of the early Holocene (Oviatt et al., 2005) when the climate in the northern hemisphere briefly cooled.

Giant Desiccation Cracks

In the Dove Creek subbasin (sec. 5, T. 10 N., R. 12 W., SLBM), giant desiccation cracks are present within the fine alluvial and lacustrine deposits of this small closed subbasin (Fig. 3.6 and Plate 3.1). There are also multiple areas of giant desiccation cracks in similar deposits of the Russian Knoll 7.5 quadrangle to the northwest. Similar structures identified as giant desiccation cracks are found in multiple other playas around the Great Basin (i.e., Neal et al., 1968; Harris, 2004; Lund et al., 2005) and are thought to be the result of the dewatering of clay rich sediment during arid conditions. The desiccation cracks exhibit a polygonal pattern with diameters from ~ 80 – 395 feet (~ 25 – 120 m). Cracks were mapped on 1:10,000 scale aerial photographs since they are difficult to discern in the field. Denser vegetation grows over the cracks, making these features noticeable on aerial photographs. Following the opening of these desiccation cracks, eolian and alluvial sediments filled the voids, trapping sediments that may have higher capillarity and provide more moisture for vegetation (Stifel, 1964).

Bedrock Stratigraphy and Geologic Structures

Limestone and quartzite deposits of the Permian and early Pennsylvanian Oquirrh basin dominate the bedrock in the quadrangle (Stifel, 1964; Doelling, 1980). The most abundant bedrock unit in the quadrangle is the Wolfcampian / Leonardian Series of the Oquirrh Group, deposited during the early Permian (Hintze, 1973). Other bedrock units in the quadrangle consist of sedimentary units deposited in the Oquirrh basin during the Permian and late Pennsylvanian, include the Rex Chert Formation, the Loray Formation, the Diamond Creek Formation, and other undifferentiated Permian and Pennsylvanian deposits (Stifel, 1964). Most of the Permian age bedrock is thicker than other areas in the Oquirrh basin and may represent a depositional low in the basin (Doelling, 1980). Deformation of the strata in the Hogup Mountains occurred during multiple phases of compression related to the Sevier orogeny during the Jurassic to late Cretaceous (DeCelles, 2004).

The area has undergone extensive Basin and Range extensional faulting from the late Tertiary to present. Normal faults bound each side of the Hogup Mountain range and unknown when these faults last ruptured (Black et al., 1999). There is evidence of Quaternary soft sediment deformation in Bonneville and Stansbury age deposits in the area; however, it is uncertain if this deformation is due to seismic activity or rapid changes in hydrostatic pressure induced by changes in the water level. The Big Pass fault has a significant scarp south of the quadrangle and may displace Bonneville age deposits at that location (Stifel, 1964). Big Pass fault probably extends north into the quadrangle from Big Pass but is concealed by lacustrine deposits. Other older small faults are present in the bedrock of the quadrangle but are not mapped.

Description of Map Units

The map can be seen in Plate 3.1 and the legend and correlation chart are presented in Fig. 3.7 and 3.8, respectively.

Alluvial deposits

Qal - alluvium (Holocene). Moderately sorted, interbedded gravels to cobbles within a matrix of sand, and silt deposited in active alluvial channels, floodplains, and minor terraces. These deposits are primarily located in the Dove Creek and Big Pass drainages. Other smaller gullies also have Qal but are not significant enough to be mapped at this scale. Maximum thickness of the deposits is less than 10 feet (3 m).

Qafy - younger alluvial fan deposits (Holocene). Post-Bonneville fans that are poorly to moderately sorted, subangular to rounded gravels to cobbles within a matrix of sand, silt, and clay. Fan deposits below the Bonneville paleoshoreline usually consist of reworked Lake Bonneville gravels and sands (Qlg), marls (Qlm), and sands (Qls) deposited by ephemeral flooding. Alluvial fan deposits originating from above the Bonneville paleoshoreline tend to be much coarser (cobble – coarse gravel) and clasts are much more angular due to the local erosion of the bedrock sources of the clasts. Deposits tend to be much coarser (cobble – coarse gravel) in the headland portion of the fans and become thinner and finer grained (medium – fine gravel) in the more distal portion of the fan slope. Expected maximum thickness of the deposits is less than 40 feet (12 m).

Qafo - older alluvial fan deposits (late Pleistocene). Pre-Bonneville or Bonneville age fans that are poorly to moderately sorted, subangular to rounded gravels to cobbles within a matrix of sand, silt, and clay. It is uncertain if deposits were active during the occupation of Lake Bonneville. Deposits tend to be much coarser (cobble – coarse gravel) in the headland portion of the fans and become thinner and finer grained (medium – fine gravel) in the more distal portion of the fan slope. These fan deposits have been extensively incised and reworked into multiple gravel (Qlg) barriers during the Provo level and the transgression to the Bonneville level. Expected maximum thickness of the deposits is less than 100 feet (30 m).

Lacustrine Deposits

Qlg - lacustrine gravel and sand (late Pleistocene). Moderately to well sorted, clast supported gravels and occasional cobbles within a matrix of coarse to fine sand. Gravel clasts are subangular to subrounded due to the proximity of bedrock sources or reworking of local alluvial fans or older lacustrine deposits. Gravel deposits were deposited during both transgressive or regressive phases of the Lake Bonneville cycle.

Well-sorted low angle (5-10º) fine grained gravel beds tend to represent shallow, lower energy beach environments and berms; coarser, moderately sorted gravels or cobbles exhibit higher angle (10-15º) beds, deposited in higher energy gravel barriers, bars, or spits. Transgressive gravel barriers tend to have high amounts of calcareous cemented gravel beds (beachrock) that dip basinward ~ 5-15º. Provo age gravel deposits tend to contain well sorted, finer grained gravels in finer grained sand matrices than transgressive gravel deposits. Thin veneers of Qlm or Qac commonly overlie transgressive gravels, whereas regressive gravels typically overlie Qlm deposits. Stansbury gravel deposits contain a high level of calcareous cement where remnants of capping tufa mounds are in float and outcrop. Regressive tufas also are prevalent on many gravel and sand (Qlg) deposits near the Provo paleoshoreline. Many gravel barriers exhibit periodic coarse to fine grained gravel interbeds and coarse to medium grained sand interbeds. Thickness of the deposits varies but ranges from thin 1-3 feet (1 m) to up to 230 feet (70 m).

Qls - lacustrine sand and silt (late Pleistocene). Moderately to well-sorted silt to fine to medium grained sand that is subrounded to rounded. Lacustrine sand deposits are interpreted as lagoonal sands or offshore sand deposits. They are rarely significant enough to be mapped or are extensively reworked by eolian processes and mapped as eolian sands (Qes); however, some significant lacustrine sand deposits are found in protected localities near the Hogup Spit (sec. 19, T. 10 N., R. 11 W., SLBM). Deposits vary in thickness and are estimated to be no thicker than 20 feet (6 m).

Qlm - lacustrine marl (late Pleistocene). Calcareous, very fine white to grey lacustrine silts and clay (marl). Marls contain various sedimentary structures and characteristics depending on their position in section. A full section of the unit typically transitions from laminated transgressive calcareous rich sand, silt, and marls on the bottom into a dense, white blocky marl topped by a clastic rich marl deposit. Ostracodes are found throughout the deposit; however, the lower laminated section is notably rich in ostracodes and occasionally very shallow ripple laminations are present. Marl laminations become thicker and denser up section, indicating a deepening lake. The blocky, dense marl contains multiple dropstones and is interpreted as the deepest deposits of the Bonneville age sediments. Dropstones, transported by ice rafting, consist of rocks from local sources (i.e., chert and limestone) or rocks (Proterozoic Elba quartzites – orthoquartzites with green, creamish white, or pink hues) from the Raft River Range to the north. The blocky marl is bounded at its upper surface by the Bonneville flood marker (BFM), representing deposition during the Bonneville flood and typically consists of a sandier unit with a layer rich in ostracodes. Locally, the layer exhibits a rusty coloration possibly due to the oxidation of the coarser sediments as groundwater preferentially flowed through the layer. Rip up clasts (2 – 4 cm in diameter) are present in the layer as marl was reworked and deposited. Up section of the BFM, marls typically become more clastic rich, indicating a regressing water level. Regressive Qlm deposits are generally thick, if they are preserved, and exhibit multiple sand deposits interbedded with calcareous silts and marls. The abundance of regressive fine grained deposits represent reworked lacustrine deposits flushed into the basin after the regression of the Bonneville flood. Many of these interbedded sand and silt deposits contain flame structures and ripple marks, and are interpreted as turbidities/storm deposits as the water level dropped. A full section of the marl can be seen in the incision of Dove Creek near the Stansbury paleoshoreline (SE ¼ sec. 20, T. 10 N., R. 12 W, SLBM), behind a substantial gravel barrier of the Stansbury paleoshoreline (NW ¼ sec. 9, T. 9 N., R. 12 W, SLBM), below the Provo paleoshorelines west of Big Pass (sec(s). 2, 10, and 11, T. 9 N., R. 12 W, SLBM), and in Peplin Flats below the Stansbury paleoshoreline (sec. 2, T. 10 N., R. 12 W, SLBM). Thickness varies from 3 to 33 feet (1 - 10 m).

Eolian Deposits

Qes - Eolian sand and silt (Holocene to late Pleistocene). Moderately to well sorted fine to medium grained sand and silt that is usually subrounded to rounded. Eolian deposits originate from lacustrine deposits, following the trace of paleoshorelines as coppice dunes deposits (Oviatt, 1991). In deeper layers of the graded sand deposits and whole gastropod samples are common. Estimated thicknesses of the deposits vary but are < 33 feet (10 m).

Playa Deposits

Qpm - playa mud (Holocene to modern). Thin clay, mud, and evaporites associated with playa and spring deposits near Dove Creek. Due to a preexisting transgressive lacustrine barrier that dammed Dove Creek, a small closed basin and lake formed at Dove Creek Springs following the regression of Lake Bonneville. This small lake overflowed the sedimentological barrier and the stream incised the barrier, eventually deepening the small internal lake and forming a small playa in the small valley. Workers from the Transcontinental Railroad (late 19th century) constructed a railroad berm across the stream valley and once again dammed the drainage to reform the small internal basin. Thickness of the deposits are less than 6 feet (< 2 m).

Fill

Qf - artificial fill (Historical). Significant fill used to provide berms for the Transcontinental Railroad. Located in the northwestern corner of the quadrangle; thickness of the deposits are from 2 to 20 feet (1-6 m).

Mixed Environment Deposits

Qac - alluvium and colluvium (Holocene and late Pleistocene). Moderately sorted to poorly sorted cobbles to gravels that are subrounded to angular. Even though alluvial and colluvial deposits are mixed, one process usually dominates the other. Colluvium dominated deposits are typically near cliffs or other steep slopes associated with relict wave cut platforms and are poorly sorted, angular gravel (pebble – cobble [4 – 256 mm]) clasts with minor portions of eolian sand and silt. Alluvium dominated deposits usually are associated with sheetwash and minor fluvial transport in small ephemeral gullies or at the base of gravel formations such as spits or bars. Alluvium dominated deposits are moderately sorted, sub angular to subrounded gravels to cobbles with a higher percentage of silts and sands. Estimated maximum thickness of the deposits is less than 20 feet (6 m).

Qla - lacustrine and alluvial deposits, undivided (Modern to late Pleistocene). Undifferentiated deposits of calcareous marl, silt, sand, and gravel consisting of lacustrine and alluvial deposits. These deposits include pre-Bonneville alluvial deposits that were reworked or etched by lacustrine processes during the occupation of Lake Bonneville or lacustrine deposits that cannot readily be determined at the scale of the map. Deposits vary in thickness and are estimated to be no thicker than 6 to 12 feet (2 to 4 m).

Stacked Unit Deposits

Qla/Bx - lacustrine and alluvial deposits over bedrock (late Pleistocene / Pennsylvanian-Permian). A thin veneer of undifferentiated lacustrine gravels, sands, silts, and marls overlying bedrock. Finer grained deposits are often slightly reworked by alluvial processes. Qla deposits are very thin (< 3 to 6 feet (1-2 m) and patchy due to the erosion of alluvial sheetwash and other rills and gullies.

Qlg/Bx - lacustrine gravel and sand over bedrock (late Pleistocene / Pennsylvanian-Permian) - Abraded bedrock platform surfaces with thin veneers of lacustrine gravels, cobbles, and sands. Lacustrine clasts are usually sub angular to subrounded due to their proximity to the bedrock sources. Capping tufa mounds and tufa encrusted gravels are often at the basinward edge of the platform. Qlg deposits are very thin, < 3 to 6 feet (~ 1-2 m).

Qlm/Qlg - lacustrine marl over gravel (late Pleistocene). Thin veneer of lacustrine marl of the transgressive and regressive phase of Lake Bonneville overlying transgressive gravel barrier ridges and bars. The dominant lacustrine morphology of these features can be delineated from general trends in aerial photos and/or determined from incisions of these features by post Bonneville ephemeral gullies. Lacustrine marls (Qlm) associated with these deposits are generally less than 3 feet (1 m) thick, and Qlg deposits can be significant (6 – 20 feet (2-7 m)) features.

Qlg/Qlm - lacustrine gravel and sand over marl (upper Pleistocene). Well to moderately sorted, clast supported gravels within a fine to medium grained sand matrix overlying transgressive and regressive marl. Gravel and sand (Qlg) deposits are associated with the regressive paleoshorelines (i.e., Provo) whereas the marls are a combination of transgressive and regressive deposits. Gravel and sand (Qlg) deposits are generally thin ranging from two 2-6 feet. (> 2 m) thick, while the marls can be up to 25 feet (8 m) thick.

Qls/Qlm - lacustrine sand and silt over marl (upper Pleistocene). Moderately to well sorted, fine to medium grained regressive sand overlying transgressive and regressive marl. These regressive sands are interpreted as either offshore sands or very shallow beach deposits. In low lying regressive deposits below the Provo paleoshoreline, very thin regressive sand deposits were usually classified as Qlg/Qlm if there was enough of a gravel component (<15% gravel). Thicknesses of the deposits vary from 6 to 14 feet (~2-4 m) near the Provo paleoshoreline and are usually < 3 feet (< 1 m) in regressive deposits below the Provo paleoshoreline.

Bedrock Units

Tb - Tertiary basalt. A grey to black aphanitic basalt flow in the Dove Creek area (sec. 5, T. 10 N., R. 12 W., SLBM). Dove Creek has dissected the larger northern portion of the basalt flow from the smaller southern lobe. The southern portion of the flow is topped by a thin layer of lacustrine marl deposited during the transgression of Lake Bonneville followed by regressive lacustrine silt/sand and reworked Aeolian sands. Basalt flows in this area have been dated at 4.3 ± 0.3 Ma via whole rock K-Ar methods (Miller et al., 1995).

Bx – pre Quaternary bedrock. Bedrock in the quadrangle is dominated by undifferentiated Permian to Pennsylvanian bedrock consisting of rhythmic layers of quartzites and chert rich carbonates. Bedrock in the quadrangle has been previously mapped in more detail; the interested reader can see Stifel (1964) or Doelling (1980) for a detailed description regarding these units.

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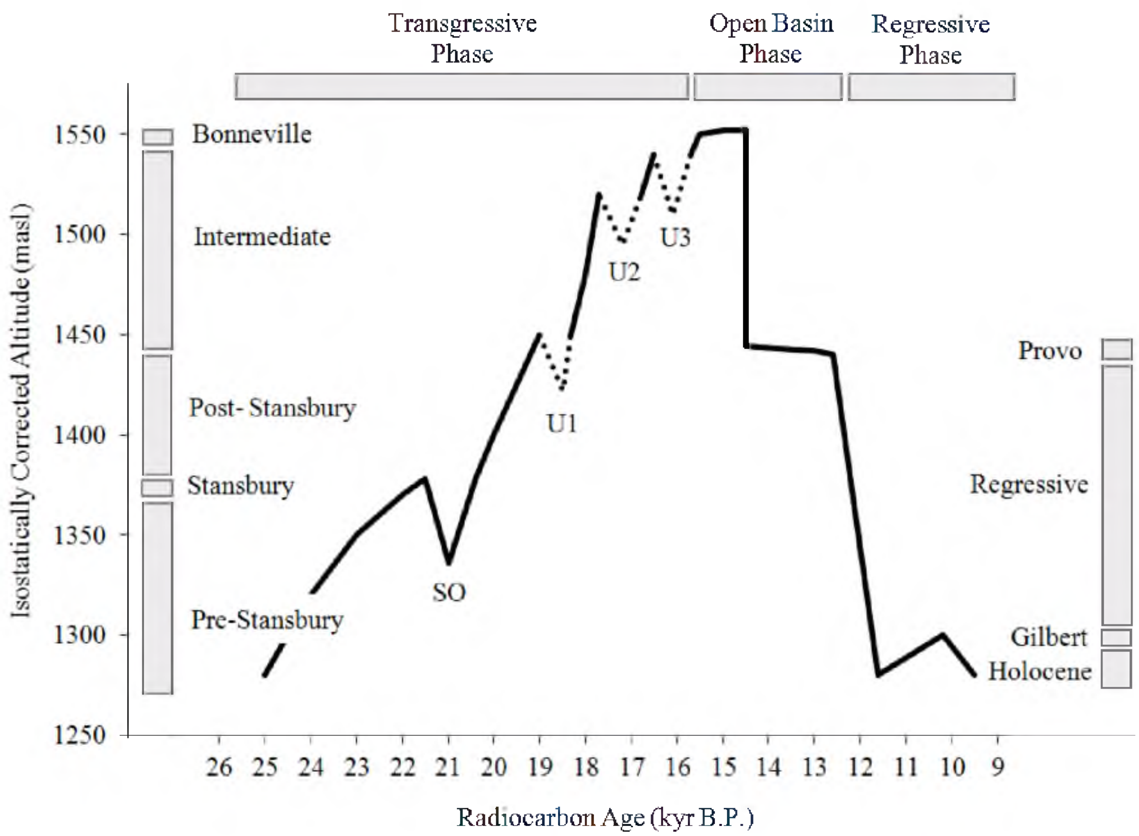


Figure 3.1: Lake Bonneville hydrograph modified from Oviatt (1997) and Godsey et al. (2011). Altitudes are adjusted for effects of differential isostatic rebound in the basin (Oviatt et al., 1992). Amplitude limits of lake stage fluctuations associated with the U1, U2, and U3 oscillations are approximate and are shown here schematically. The temporal range of the transgressive, regressive, and open phases of the lake cycle are shown horizontally, whereas the altitudinal range of each of the paleoshorelines is shown vertically within either the transgressive or the regressive phases of the lake cycle.

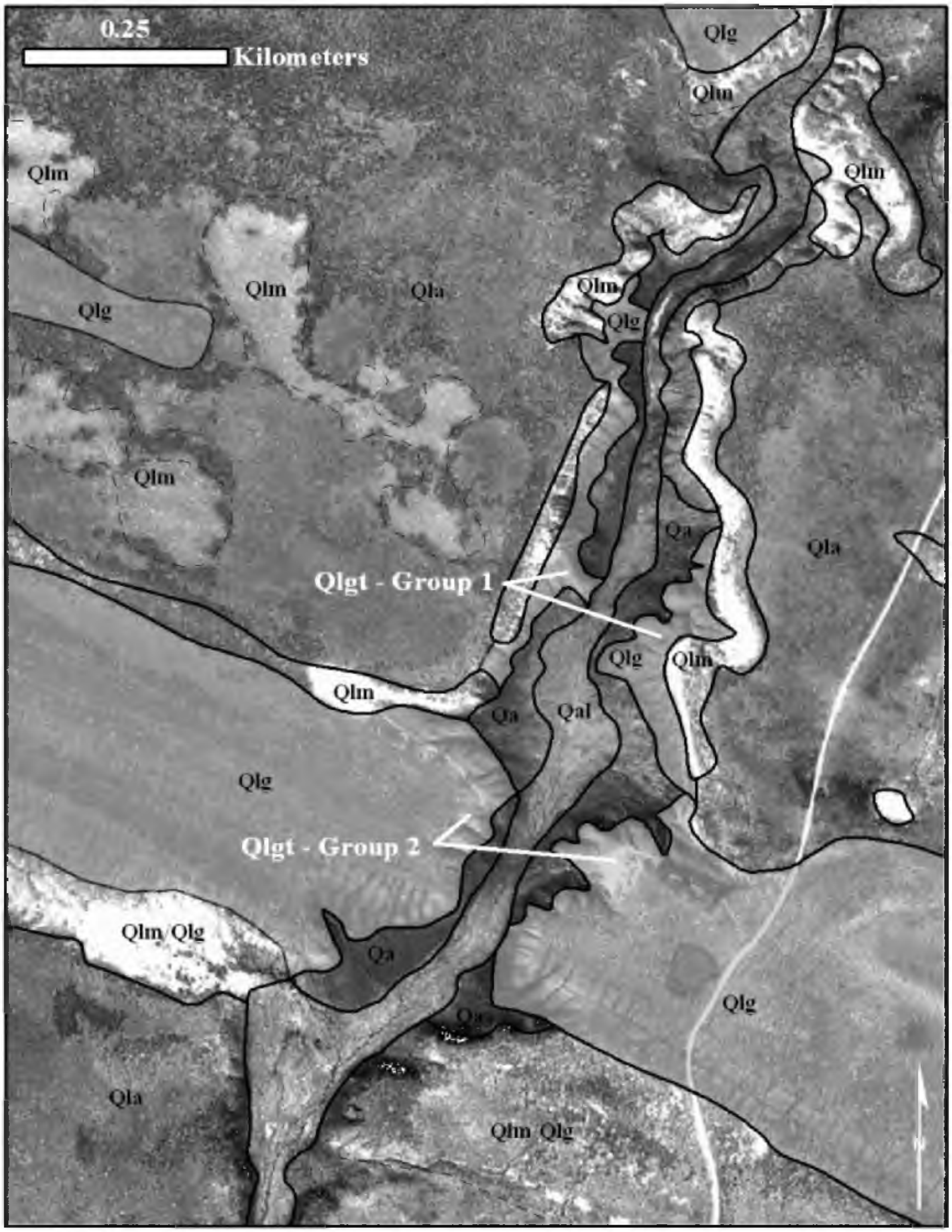


Figure 3.2: The Stansbury shorezone is incised by Dove Creek to expose multiple gravel ridges overlain with fine grained lagoonal sands and silts (Qll), transgressive deepwater marls (Qlmt), and regressive marls and sands (Qlmr). (sec(s). 20 and 21, T. 10 N., R. 12 W., SLBM)

Figure 3.3: The embayment south of the Hogup Spit and the relationship of the Bonneville and the Provo levels with multiple Intermediate barrier ridges. Photo taken from (NE ¼ sec. 25, T. 10 N, R. 12 W., SLBM) looking to the northeast. Truck and height of spit can be seen for scale.

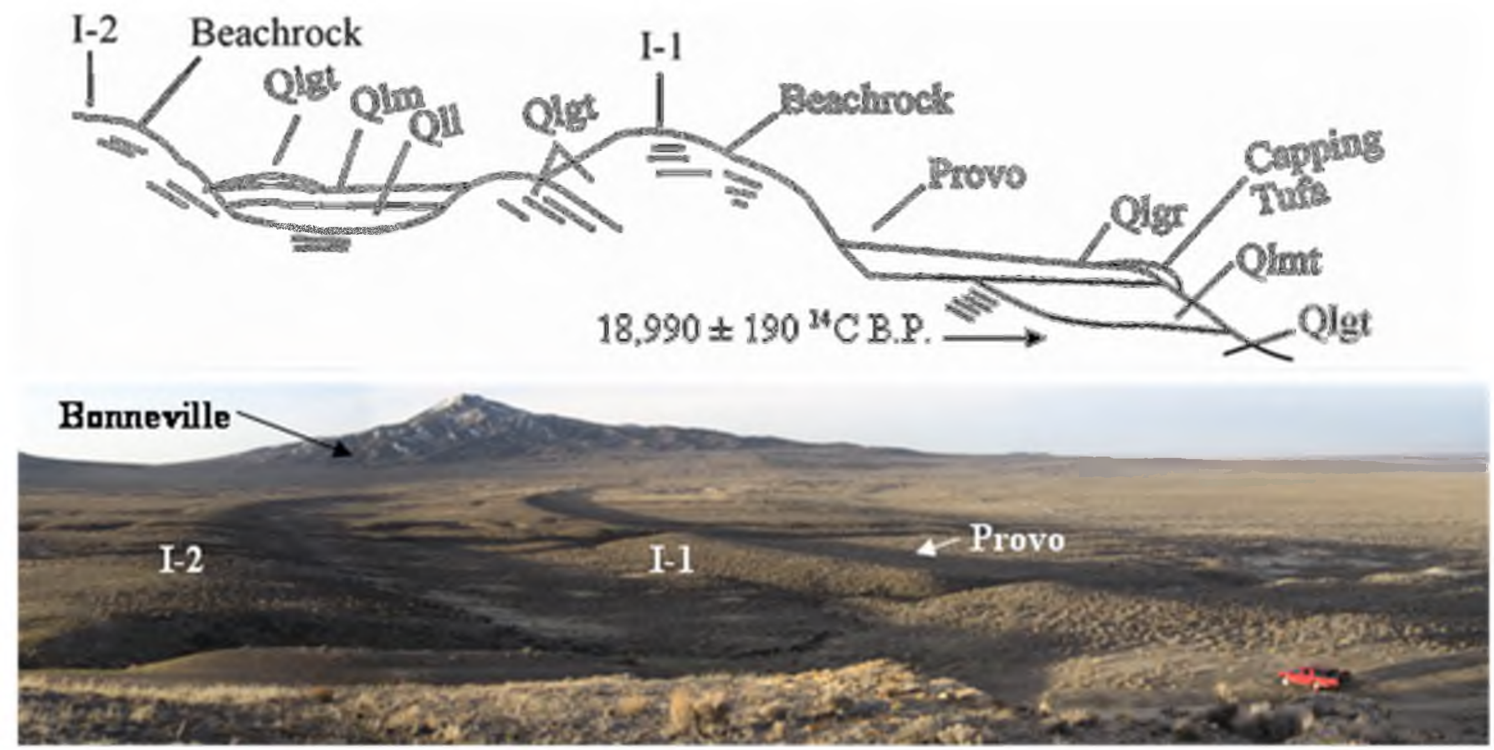


Figure 3.4: Relationship between I-1, I-2, and Provo deposits at Big Pass. Qlgt: transgressive age lacustrine gravels, Qlmt: transgressive age fine grain lacustrine sediments (marls and calcareous rich sand, silt and clay), Qll: lacustrine lagoonal muds and sands, Qlgr: regressive lacustrine gravels (Provo age), Beachrock: carbonate cemented gravels, and capping tufa deposits. Radiocarbon sample was measured from gastropod sample collected from transgressive offshore lacustrine sands (Sack, 1999). Red truck for scale. Photo taken from (SE ¼ sec. 35, T. 10 N, R. 12 W., SLBM) looking to the south. Radiocarbon age of intermediate gravels underlying Provo deposits have been dated at 18,990 ± 190 ^{14}C B.P. Truck for scale.

Figure 3.5: View of Provo age deposits overlying transgressive gravel and sand (Qlgt) and marl (Qlmt), south of Big Pass(NW ¼ sec. 15, T. 9 N, R. 12 W., SLBM). Photo taken from the east of the bar, height of bar for scale

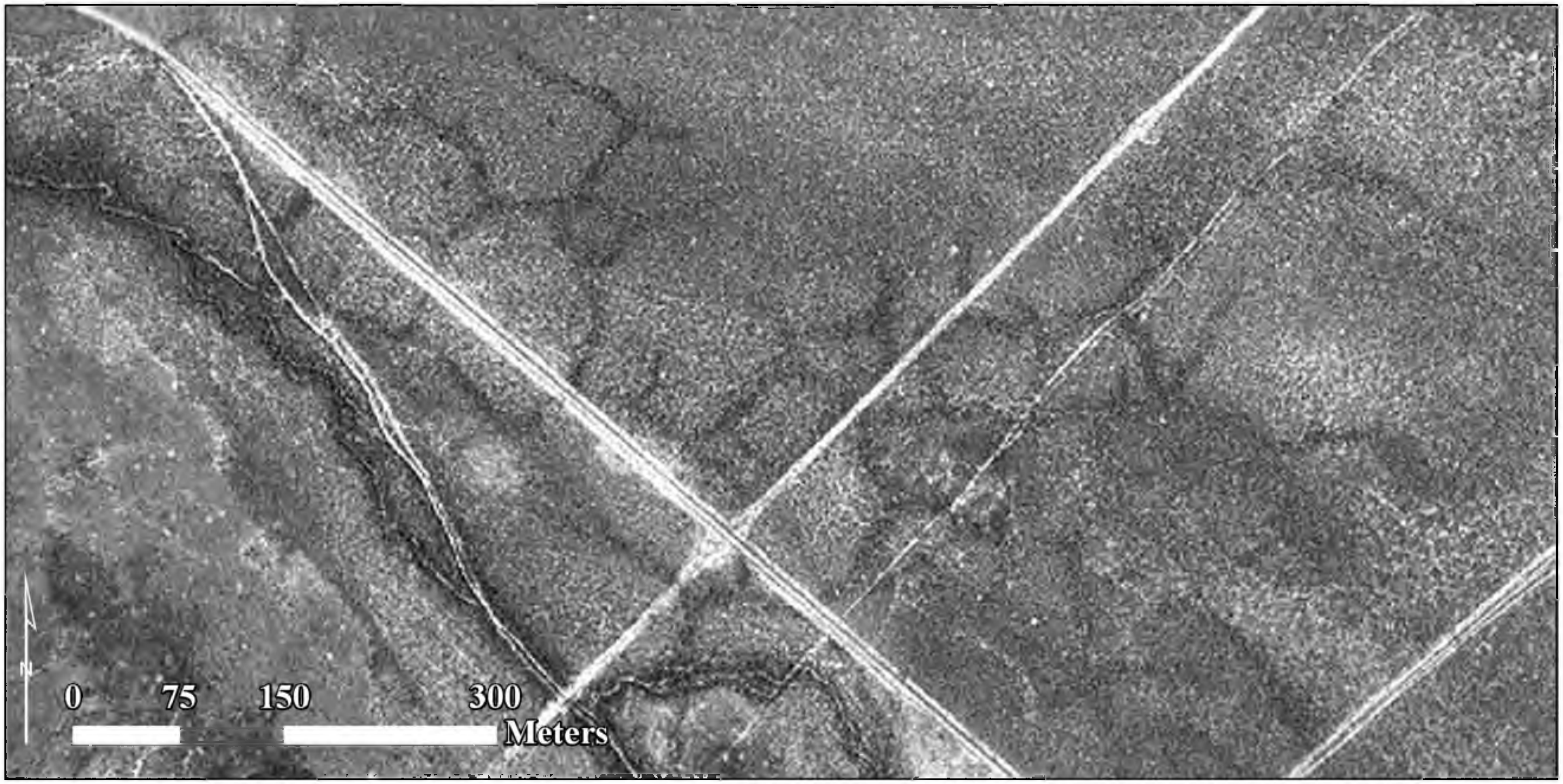


Figure 3.6: Dove Creek subbasin desiccation cracks (sec. 5, T. 10 N., R. 12 W., SLBM)

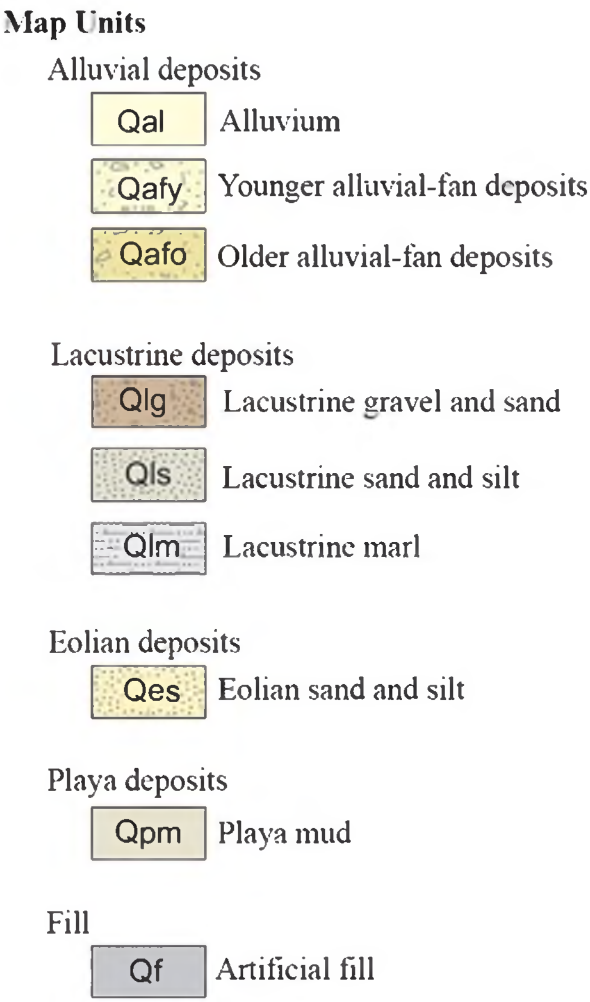


Figure 3.7: Legend for Plate 3.1

Mixed-environment deposits

Qac	Alluvium and colluvium
Qla	Lacustrine and alluvial deposits, undivided

Stacked-unit deposits

Qla/Bx	Lacustrine and alluvial deposits over bedrock
Qlg/Bx	Lacustrine gravel and sand over bedrock
Qlm/Qlg	Lacustrine marl over gravel
Qlg/Qlm	Lacustrine gravel and sand over marl
Qls/Qlm	Lacustrine sand and silt over marl

Bedrock Units

Tb	Tertiary basalt
Bx	Pre-quaternary bedrock

Geologic Symbols

^{14}C-1 Radiocarbon sample location and number

Well

Sand and gravel pit

Contact - Dashed where approximately located; dotted where uncertain

Paleoshoreline of the Bonneville lake cycle - Mapped at the wave-cut bench of erosional paleoshorelines or at the crest of constructional beach barriers, bars, and spits. Paleoshorelines are color coded based on age/identity.

Paleoshoreline - Dashed where uncertain age/identity; dotted where concealed

Crest of barrier ridge, bar, or spit - Dashed where uncertain age/identity

Transgressive paleoshorelines

S Stansbury paleoshoreline

B Bonneville paleoshoreline

Y Other transgressive paleoshoreline

Regressive paleoshorelines

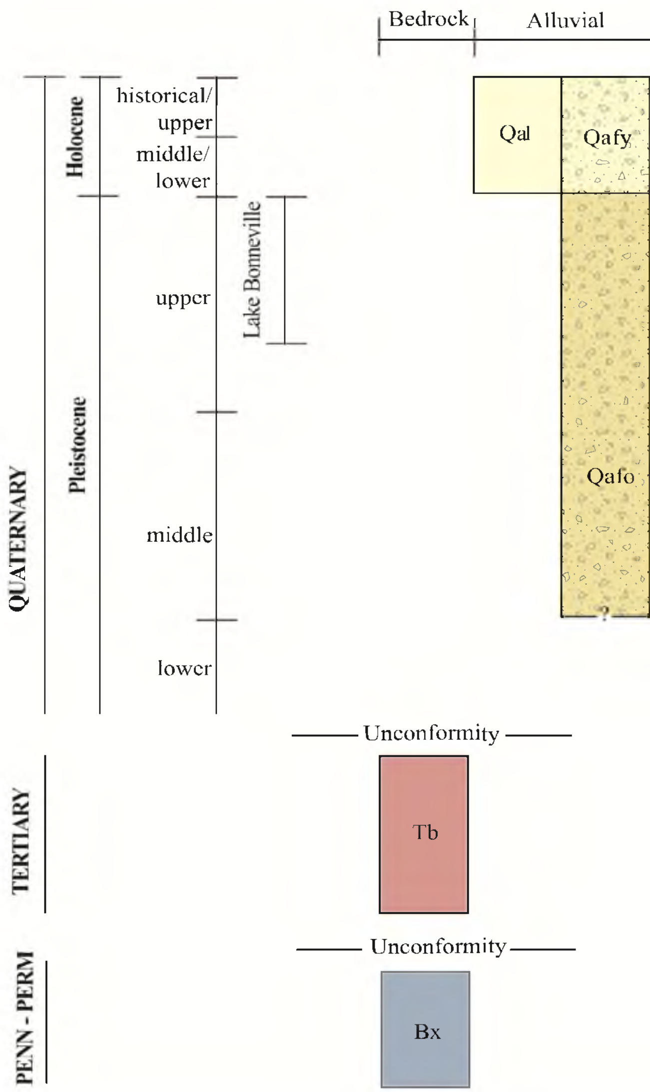
P Provo paleoshoreline

G Gilbert paleoshoreline

X Other Regressive paleoshoreline

Figure 3.7 (cont.)

Figure 3.8: Correlation chart of geologic units for Plate 3.1



Bedrock
Alluvial
QUATERNARY
Holocene
Pleistocene
historical/ upper
middle/ lower
upper
middle
lower
Lake Bonneville
Qal
Qafy
Qafo
Unconformity
TERTIARY
Tb
Unconformity
PENN - PERM
Bx

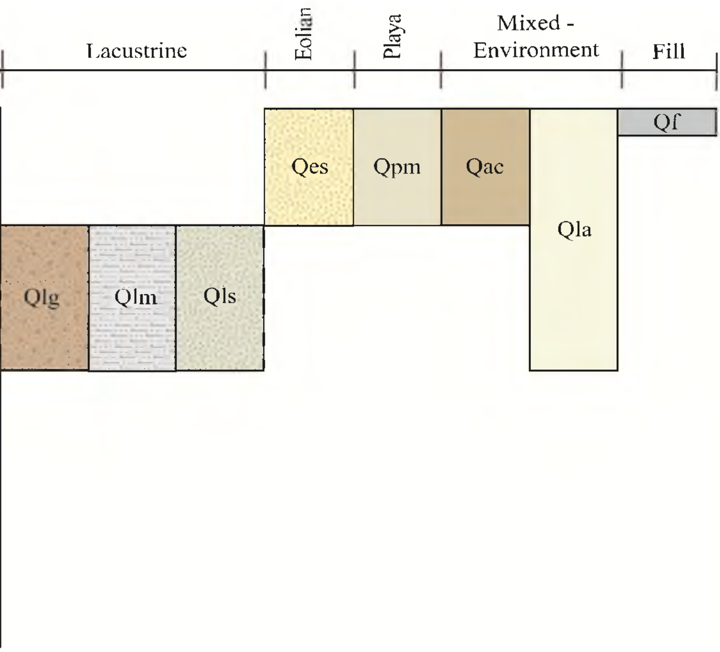

Lacustrine
Eolian
Playa
Mixed - Environment
Fill
Qes
Qpm
Qac
Qla
Qf
Qlg
Qlm
Qls

Table 3.1: Lake Bonneville paleoshoreline ages and altitudes.

Age ranges and general altitudes of Lake Bonneville paleoshorelines within the Hogup Bar quadrangle.

Lake Phase	Paleoshoreline (map symbol)	^{14}C yr B.P.	Cal. Yr B.P.[1]	Local Altitude (meters)	Isostatically Corrected Altitude[2] (meters)
Transgressive	Pre-Stansbury	~25,000-22,000[3]	29,000-23,200	~1338-1357	~1320-1336
	Stansbury (S)	22,000 – 21,000[4]	24,000-23,200	1360-1405	1339-1378
	Post-Stansbury	21,000~19,000	23,200-19,200	~1405-1480	~1378-1443
	Intermediate[6]	~19,000-15,300	20,400-18,300	~1487-1598	~1449-1545
	Bonneville (B)	15,300 – 14,500[7]	18,300-17,400	1598-1608	1545-1,552
Regressive	Provo (P)	14,500 – 12,600[8]	17,400-15,000	1458-1483	1424-1445
	Regressive[9]	~12,600 – 11,000	15,800-10,800	~1310-1458	~1295-1424
	Gilbert (G)	11,000 – 10,000	10,800-9,300	1305-1310	1291-1296

[1]Calendar calibrated ages of most paleoshorelines have not been published. Calendar calibrated ages shown here have been estimated by using Calib 6.0 (Stuvier & Reimer, 1993).

[2]Paleoshoreline elevations were corrected by using the methodology of Currey and Oviatt (1985).

[3]Estimated based on altitude in comparison to the lakes hydrograph published by Oviatt (1997). – Dates have not been calibrated past 25,000 calendar yrs B.P.

[4]Oviatt et al. (1990). Currey (written communication to the Utah Geological Survey, 1996) assumed a maximum age for the Stansbury paleoshoreline of 21,000 ^{14}C yr B.P., which is used in the conversion to calendar yrs.

[5]Lowr intermediate paleoshorelines are transgressive paleoshoreline features positioned in between the altitudinal limits of the Provo and Stansbury paleoshorelines. The extent of the age of the paleoshoreline features is based on the extent of radiocarbon dates summarized in Sack (1999) and Oviatt (1990).

[6]Upper intermediate paleoshorelines are transgressive paleoshoreline features positioned in between the altitudinal limits of the Provo and Bonneville paleoshorelines. The extent of the age of the paleoshoreline features is based on the extent of radiocarbon dates summarized in Sack (1999) and Oviatt (1990).

[7]Oviatt et al., (1992), Oviatt (1997)

[8]Godsey et al., (2011) revised the timing of the occupation of the Provo paleoshoreline and subsequent regression.

[9]Regressive paleoshorelines positioned in between the Gilbert and Provo levels of the lake system. The estimated age and altitude range of these paleoshorelines are based on the constraints of the Provo regression and the altitudinal limits of the Gilbert paleoshorelines.

Table 3.2: Radiocarbon ages of samples from the Hogup Bar quadrangle

Num	Lab Num.[1]	Lat.	Long.	Altitude [2]	Method [3]	Dated Material	Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal. BP)[4]	Depositional Setting	Reference
1	Beta-57132	41.539	113.170	1,556 (1,512.5)	radiometric	mollusk shells	18,990 ± 190	22,960 (22,530) 22,130	Top of transgressive gravel underlying transgressive fine sand	Sack, 1999
2	Beta-307248	41.588	113.138	1,505 (1,467.7)	AMS	*Stagnicola*	18,510 ± 70	22,290 (22,180) 22,050	Fine sand above beach gravels that gradates into a sandy marl	this study
3	Beta-307247	41.588	113.139	1,517 (1,478.3)	AMS	*Stagnicola*	18,240 ± 70	22,060 (21,590) 21,510	Fine sand above beach gravels that gradates into a sandy marl	this study
4	Beta-307251	41.589	113.139	1,526 (1,486.2)	AMS	*Stagnicola*	17,210 ± 70	20,490 (20,370) 20,290	Fine sand above silt/sand rich marls underlying beach gravels	this study
5	Beta-307250	41.588	113.138	1,525 (1,485.3)	AMS	*Stagnicola*	16,930 ± 60	20,270 (20,170) 20,040	Fine sand above beach gravels that gradates into a sandy marl	this study
6	Beta-246724	41.577	113.141	1,512 (1,473.9)	AMS	*Stagnicola*	16,910 ± 80	20,280 (20,030) 19,860	Within slightly cemented intermediate gravel barrier	this study
7	Beta-246723	41.578	113.141	1,556 (1,512.5)	AMS	*Stagnicola*	16,770 ± 70	20,120 – 19,800 (19,910) 19,640 – 19,630	Fine sand lagoon deposits behind intermediate gravel barrier	this study
8	Beta-307249	41.589	113.139	1,507 (1,469.5)	AMS	*Stagnicola*	16,480 ± 70	19,810 – 19,590 (19,550) 19,700 – 19,440	Fine sand above clay/silt rich marls underlying beach gravels	this study

Table 3.2: *(continued)* Radiocarbon ages of samples from the Hogup Bar quadrangle

Num	Lab Num.[1]	Lat.	Long.	Altitude [2]	Method [3]	Dated Material	Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal. BP)[4]	Depositional Setting	Reference
9	Beta-246725	41.507	113.217	1,470 (1,437)	AMS	*Stagnicola*	16,280 ± 60	19,530 (19,460) 19,280	Gravel beach barrier - reverse grading deposit from coarse gravels and sand overlying finer sands	this study
10	Beta-169097	41.594	113.183	1,467 (1,434.4)	AMS	*Stagnicola*	12,430 ± 50	15,412 ----------- 14,144	Fine sand overlain by sandy gravel in embayment between spit & main Provo Shoreline	Godsey et al., 2005
11	Beta-169098	41.593	113.186	1,465 (1,432.6)	AMS	*Stagnicola*	11,910 ± 50	15,195 ----------- 13,621	Fine sand grading upward into sandy gravel at toe of spit, lies on top of gravel bench	Godsey et al., 2005

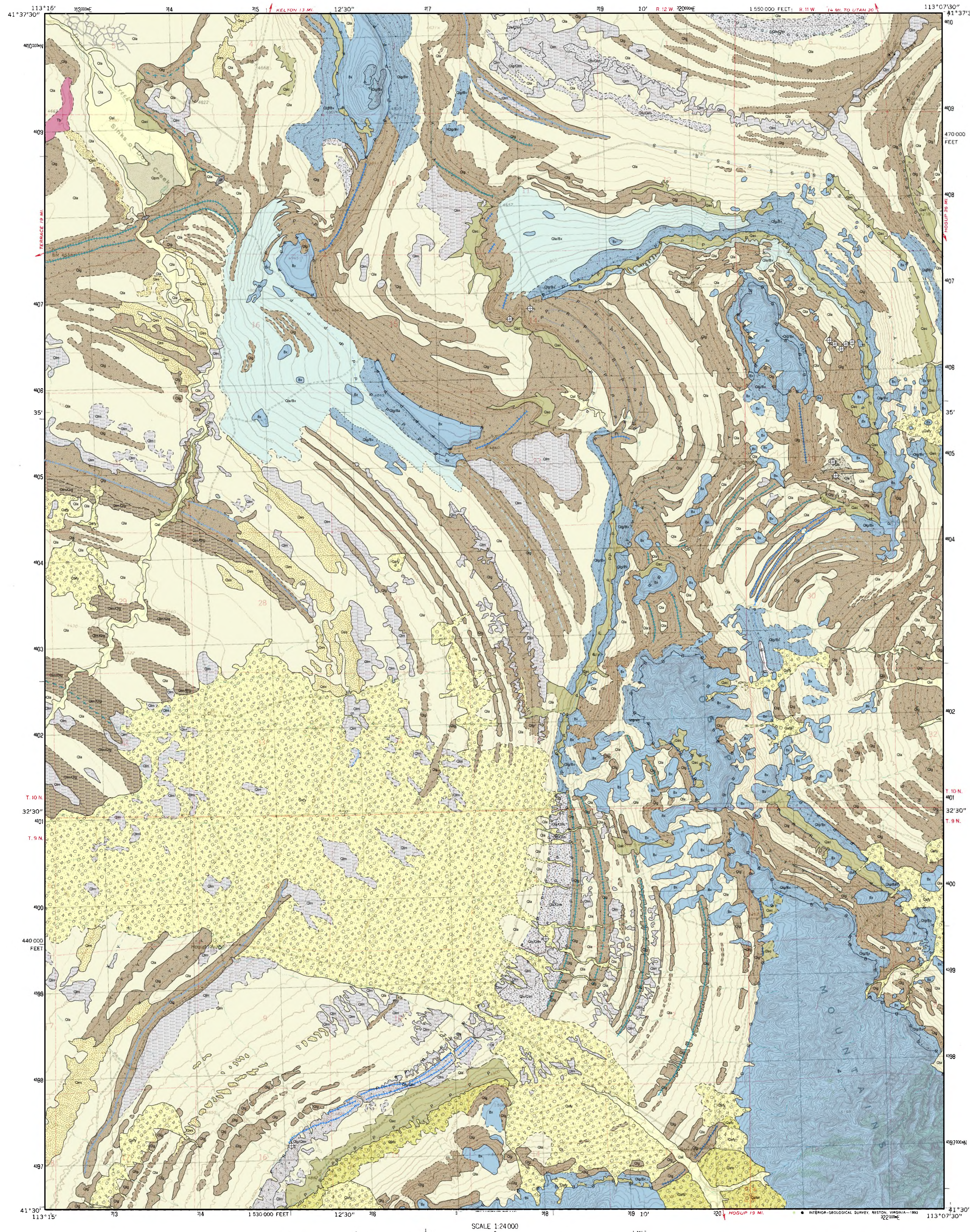
1. Radiocarbon ages by Beta Analytic Lab, Miami, Florida.

2. Altitude in meters above sea level, values in parentheses are corrected for isostatic rebound (Oviatt et al., 1992)

3. AMS (Accelerator Mass Spectrometry)

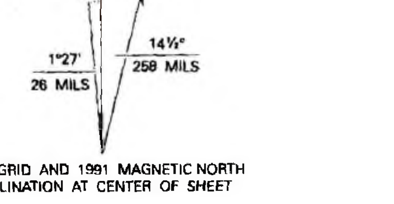
4. Value in parentheses is calibrated age, and values outside of the parentheses are the maximum and minimum ages of the sample to two sigma. Calibrated dates are reported or estimated from CALIB 4.0 program (Stuiver and Reimer, 1993).

Plate 3.1: Geologic map of the unconsolidated deposits in the Hogup Bar quadrangle, Box Elder County, Utah



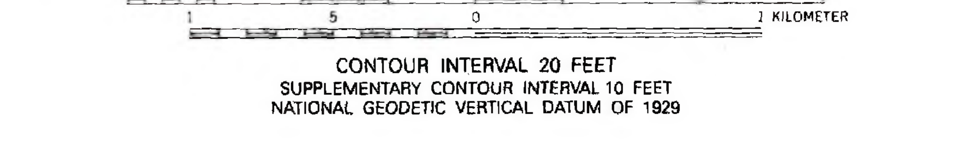
Base map from U.S. Geological Survey
Hogup Bar quadrangle, 1991
Geologic data and base map in NAD 1927

This geological map was funded by the U.S. Geological Survey, National Cooperative Geologic Mapping Program through USGS EDMAP award number 08HQAG0070. The views and conclusions within this document are those of the authors and do not necessarily represent those of the Utah Geological Survey.

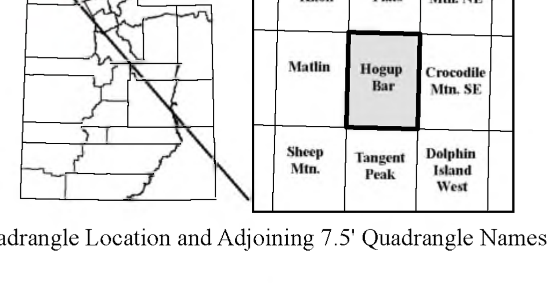


UTM GRID AND 1991 MAGNETIC NORTH
DECLINATION AT CENTER OF SHEET

SCALE 1:24 000



CONTOUR INTERVAL 20 FEET
SUPPLEMENTARY CONTOUR INTERVAL 10 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929



Quadrangle Location and Adjoining 7.5' Quadrangle Names

Plate 3.1

GEOLOGIC MAP OF THE UNCONSOLIDATED DEPOSITS IN THE HOGUP BAR QUADRANGLE, BOX ELDER COUNTY, UTAH

by
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2012

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CHAPTER 4

CORRELATION OF THE LATE PLEISTOCENE "INTERMEADIATE" PALEOSHORELINES OF LAKE BONNEVILLE, U.S.A

Abstract

Multiple transgressive paleoshorelines of the late Pleistocene Lake Bonneville have been identified between the altitudinal limits of the Bonneville and Provo levels and were named by G.K. Gilbert as the "Intermediate" paleoshorelines. The chronologic record of these Intermediate paleoshorelines is relatively unknown because they have not been correlated in the basin due to their individual altitudinal variations. The complexity of the altitudinal variations associated with the Intermediate paleoshorelines has been attributed to differential hydro isostatic rebound patterns, differential wave energy patterns, and autogenic sedimentological processes related to an oscillating lake. In order to correlate the features locally and basin wide, modern GIS technologies are coupled with field investigations of the sedimentological and geomorphic patterns of the features. High resolution digital elevation models were utilized in an new hydrostatic rebound model and then a potential wave energy model was developed by incorporating fetch and slope as proxies for wave energy. Both of these models, coupled with field investigations, are used to quantify factors that influence the development of paleoshorelines in a

complex lacustrine system. Six significant and multiple small Intermediate paleoshorelines are correlated in the northwestern portion of the basin (Hogup Mountains and Matlin Mountain) and within a southern arm of the basin (Wah Wah valley). Correlation of these paleoshorelines allows a more accurate history of the paleoshorelines and the basins climatic history to be determined. Correlating these Intermediate paleoshorelines also have the potential to further aid in the understanding of how the lake may have responded to regional and/or global submillennial climatic shifts and how the basin may respond to these same type of climatic shifts in the future.

Introduction

G.K. Gilbert is known as one of America's great geomorphologists (Pyne; 1980; Burstyn, 1984; Sack, 1991). His work ranged from coastal (Gilbert, 1890), glacial (Gilbert, 1903), volcanic (Gilbert, 1877), tectonic (Gilbert, 1906; 1909), or even extraterrestrial geomorphology (1893). However, he is most known for his monumental work regarding the late Pleistocene Lake Bonneville (1890). G.K. Gilbert's (1890) research regarding Lake Bonneville laid the foundation for multiple studies that investigated a variety of geodynamic, geomorphic, paleontological, sedimentological, and anthropologic processes regarding the lake (Oviatt and Thompson, 2002).

Gilbert (1890) discusses a subset of minor paleoshorelines that he termed the "Intermediate" paleoshorelines (defined as paleoshorelines between the altitudinal limits of the Bonneville and Provo lake levels). Gilbert identified multiple localities in the basin where these Intermediate paleoshorelines were prevalent (e.g., Wah Wah Valley, the Old River Bed, Stockton Bar, Dove Creek, Grantsville spits—Fig. 4.1). He observed that at each locality, the crests of these Intermediate features were not at consistent altitudes and/or the entire suite of paleoshorelines was not all preserved; therefore, Gilbert was unable to determine how the paleoshorelines related to the overall history of the lake. Even though other researchers have briefly discussed these less prominent features (Gilbert, 1890; Scott et al., 1983; Scott, 1988; Burr and Currey, 1988; Oviatt et al., 1994; Sack, 1999) and periodically included them in regional geologic maps (Miller & Oviatt, 1994; Miller & McCarthy, 2002), the literature does not expand on how the variability of these paleoshorelines relate to the chronologic record of the lake and regional climatic history.

One of the hypotheses of this study is that the intermediate paleoshorelines record regional and/or global submillennial climatic shifts. The successor of Lake Bonneville, the modern Great Salt Lake, has fluctuated by up to 6 m over the last century and caused dramatic changes in the extent of the lake and flooding hazards in the region (Atwood, 1994; 2006). Without understanding the longer term responses of lake levels, how the lake may respond to future climatic shifts cannot be predicted. We propose that Intermediate paleoshoreline features can be correlated by quantifying their altitudinal extents and by determining causes for variations in their altitude. This will then lead to a better chronological record for the transgressive period of the lake's history, how the lake responded to past climatic shifts, and how it may respond to future climatic shifts.

Gilbert's original hypothesis was that the Intermediate paleoshorelines could be correlated around the basin by measuring their crest height. The crest height could then be analyzed to determine the chronology of individual lake stage heights and aid in the history of the region's water budget and climate. However, he soon found that it was very difficult to correlate these Intermediate features around the basin due to the inconsistency of paleoshoreline altitudes and the variability of how paleoshorelines were developed and preserved in the geologic record. Gilbert arrived at three hypotheses to explain the dominant cause of altitudinal variation differences of paleoshorelines: 1) variation due to differential hydro isostatic load of the different depositional horizons, 2) variation due to depositional patterns induced by wind and waves during storm events, and 3) variation due to the autogenic (internally driven) sedimentological processes involved with an oscillating water level. Gilbert concluded that hydro isostatic rebound could not be the principal cause for the local variation of these Intermediate paleoshorelines. The difference between the altitude of the Bonneville and Provo paleoshorelines is consistent within localized regions, yet the altitude differences between individual Intermediate paleoshorelines is not consistent within the same regions. Gilbert observed that the altitude of the erosional features (i.e., erosional notches and wave platforms) of an individual paleoshoreline differed from the attitude of depositional landforms (i.e., barrier and spit crests). He suggested that depositional landforms would overestimate the still water level (SWL) due to the effects of storm wave run up, whereas erosional platforms better estimated the SWL. Gilbert suggested that differential wave energy could cause significant variation in the altitude of the Intermediate paleoshorelines; however, he believed that this phenomenon was not the main cause of the variations in altitude. Gilbert based this interpretation on the fact that he did not see strong sedimentological evidence of coarser material on the higher bars within a set of paleoshorelines. He suggested that the grain size of the material was relatively homogeneous within the various bars; therefore, it could be inferred that the prominent difference in their altitudes was not due to the differential deposition patterns of features caused by stronger wave energy. Gilbert determined that the main cause for the altitudinal variations of the Intermediate paleoshorelines was due to heterogeneities of the natural autogenic sedimentological processes (i.e., change in distribution and abundance of sediment, heterogeneity of wave energy distribution, changes in accommodation space, and changes in topography of bedrock and landforms) that occur throughout the basin during a transgressing and oscillating lake level. As the water level in the lake oscillated, the sediment supply, wave conditions, and morphology of the shorezone changed from region to region (i.e., Carter and Orford, 1993; Forbes et al., 1995; Catteneo and Steel; 2003; Klinger et al., 2003; McMillian and Teller, 2012; Tamura, 2012). The wave energy pertaining to a specific level also changed and caused the sediments from preexisting features to be reworked and/or buried (i.e., Orford et al., 1996; Houser and Greenwood, 2005). Paleoshorelines may be preserved in one location but then eroded away or buried in another due to changes in sediment sources, landform/bedrock morphology, or wave distribution (i.e., Carter and Orford, 1993; Forbes et al., 1995; Tamura, 2012).

This study correlated the Intermediate paleoshorelines in the basin by using high resolution digital elevation models (DEM's) and GIS technologies coupled with more traditional sedimentological and geomorphic field studies. The study investigates the hypotheses suggested by Gilbert (1890) to see how these factors influenced the variations in the altitude of the Intermediate paleoshorelines. It also discusses a model to provide a first approximation of how the Intermediate paleoshorelines are correlated by correcting for isostatic rebound and by defining regional differences found in the wave energy and sedimentological factors.

Geologic Setting

The Bonneville lake cycle was initiated approximately 27,000 ^{14}C yr before present (B.P.) and ended approximately 10,000 ^{14}C yr B.P. at the end of the Last Glacial Maximum (Oviatt, 1997). Continual changes in the lake's water budget and resultant water level have provided a substantial repository of sedimentological and geomorphic paleoshoreline features that record the region's climatic history. The general hydrologic chronology and the broad climatic history inferred by the four major paleoshorelines of the lake cycle (i.e., Stansbury, Bonneville, Provo, and Gilbert lake levels; Fig. 4.2) have been well established (Currey and Oviatt, 1985; Oviatt et al., 1992; Oviatt, 1997; Kaufman, 2003; Balch, 2005; Godsey et al., 2005; Oviatt et al., 2005). Currey and Oviatt (1985) and Oviatt (1997) suggest that multiple large (20 – 45 m) oscillations occurred during the transgression of the lake. The Stansbury Oscillation(s) (~20,000–22,000 ^{14}C yr B.P.) is the most prominent and well studied of these oscillations and may be composed of two separate oscillations (Oviatt et al., 1990; Patrickson et al., 2010). Oviatt (1997) suggested up to three more large transgressive oscillations (termed U1–U3) and Nelson and Jewell (in review, Chap. 5) suggested two more oscillations from 16,000 to 23,000 ^{14}C yr B.P. Some of these oscillations have been tentatively correlated to global climatic events such as Heinrich events (Oviatt, 1997).

Lake Bonneville rose to its maximum altitudinal limit (~1,552 masl) ~15,500 ^{14}C yr B.P. that roughly correlates to the timing of the local glacial maximum of glaciers within the Bonneville basin (Refsnider et al., 2008). Once the lake level reached the altitude of a threshold near Zenda, Idaho, the lake overflowed into the Snake River/Columbia River Drainage Basin. When the lake reached this threshold, the water level stabilized for ~1,000 yrs to form the paleoshorelines of the Bonneville level. Consistent isostatic adjustment of the region depressed the basin floor during the duration of the lake at the Bonneville level and caused some Bonneville paleoshorelines to have multiple altitudinal expressions (Gilbert, 1890; Burr and Currey, 1988). The threshold at Zenda was composed of the weakly consolidated Salt Lake Formation and other unconsolidated alluvial deposits (Janecke and Oaks, 2011). Approximately 14,500 ^{14}C yr B.P. (O'Conner, 1993) the lake catastrophically breached the threshold, dropping its water level by ~108 m and losing ~5,238 km^3 of water. The flow of the flood caused massive erosion downstream as the water moved down the Snake River and Columbia River drainages to the Pacific Ocean. Following the flood, the lake was constrained by the continued overflow of a new topographical divide south of Red Rock Pass, Idaho (Janecke and Oaks, 2011). Due to lake overflow at this new threshold, the water level remained relatively constant. Substantial paleoshorelines, known as the Provo paleoshorelines, developed as the lake remained at this threshold until ~12,500 ^{14}C yr B.P. (Godsey et al., 2011). However, just like the Bonneville level, the basin isostatically adjusted to the new volume of water at the Provo level, causing the expressions of the Provo paleoshorelines to be represented at a range of altitudes (Burr and Currey, 1988; Godsey et al., 2005).

Following the occupation of the lake at the Provo level, the water level quickly fell below historic levels for the modern Great Salt Lake (Benson et al., 1992). This dramatic regression is hypothesized to be related to the Bølling Allerød interstadial (Benson et al., 2011; Godsey et al., 2011) and thought to have lasted ~500–1,000 yrs. The lake then transgressed briefly in the early Holocene to form the Gilbert highstand that may be related to the cold period of the Younger Dryas stadial (Oviatt et al., 2005).

Determining the Still Water Level

The still water level (SWL) is the altitude of a water surface in the absence of waves. In terminal lake basins, the SWL will fluctuate more than in open systems. For example, over the last 150 yrs the SWL of the modern Great Salt Lake fluctuates an average of ~0.5 m per yr over a 6 m interval (Atwood, 1994). A good estimate of a SWL acts as a reference for any given lake surface to estimate the amount of lake setup and storm run up that a paleoshoreline may record. In a paleolake, various authors suggest that the SWL is best represented in the record as the hinge line (the point where the top sets of the features steepen into the backsets) of a depositional feature (Adams and Wesonsky, 1998) or the erosional notch of a wave platform (Gilbert, 1890; Trenhaile, 1987; Currey and Sack, 2009). The hinge line can also be estimated by looking at the inflection point between the convex to concave profile of the depositional landform. However, these geomorphic features can easily be buried by post processes of colluvium or alluvium and they are not always readily visible in the geomorphic record.

Determining the SWL at any given point requires some interpretation of the potential wave energy that reached that site. The altitudes for these features are either measured in the field via differential GPS or modeled from a 5 m resolution digital elevation model (DEM) procured from the Utah Automated Geographic Reference Center. The altitudes of depositional landform crests and unburied portions of erosional platforms are prevalent in the geomorphic record and are often used as an estimate for the

SWL. However, the crestal altitude of depositional landforms typically overestimates the SWL, and the average altitude of an erosional platform usually slightly underestimates the SWL (Sunamura,1992; Trenhaile, 1987; 2010; 2011); (Fig. 4.3). For example, the Hogup Spit is ~8 m lower than the inflection point of the erosional platform that is less than 100 meters away. The erosional platform is probably a better representation of the SWL, so at this locality, it is estimated that the SWL for the Bonneville is ~1593 masl (Table 4.1). If these geomorphic features are not available, the SWL is estimated by the altitudinal crest average of the depositional landforms within a set of paleoshorelines. This average crest altitude will be a slight over approximation of the SWL and the error will vary depending on the wave energy, sedimentological, and geomorphic conditions for the location. In this study, a hierarchal order of estimating the SWL: 1) erosional notches of a paleoshoreline, 2) the wave platform surface, 3) the hinge line of a depositional feature, and 4) the average altitudinal crest of a depositional feature (Fig. 4.4).

Hydro Isostatic Correction Models

The hydro isostatic rebound in the basin needs to be quantified in order to assess Gilbert's hypothesis regarding its control on the Intermediate paleoshorelines. Past researchers have used two basic methodologies to determine the isostatic rebound of the basin (i.e., Oviatt et al., 1992: Bills et al., 2001). These two methodologies were tested and then modified to compare and contrast their ability to correct for the hydro isostatic rebound.

The first correction uses the methodology of Bills et al. (2001) and will be referred to as the Isostatic Surface Model (ISM). For the ISM, Bills et al. (2001) created a 90 m resolution isostatically corrected DEM for both the Bonneville and Provo levels. To check the 90 m resolution model for accuracy, a contour of the predicted lake extent was extracted from the corrected DEM at the Bonneville and Provo levels and compared to preexisting mapped paleoshorelines and paleoshoreline features seen in aerial photography. The 90 m model resolution did not accurately predict the lateral extent of the lake at these levels since paleoshoreline features were off by an average of 40 m. Five m resolution DEMs are now available for the basin so the ISM was updated by adapting the methodology of Bills et al. (2001) using these higher resolution DEMs.

The ISM used the paleoshoreline dataset of Currey (1982) that includes the latitude, longitude, and altitude of paleoshoreline points around the basin of the Bonneville and Provo levels. To determine the validity of the isostatic correction of the ISM, it is necessary to discuss the potential errors with Currey's (1982) dataset used to create the interpolated surface.

1) The coordinates and altitudes of each point were extracted from topographical maps, GPS, or surveys of the sites from known altitude benchmarks; therefore, depending upon the method used, the accuracy of site coordinates and altitude can vary.

2) The altitude of a paleoshoreline does not necessarily represent a SWL. The crest of a depositional landform usually represents the maximum extent of storm deposits (Gilbert, 1890; Currey, 1982; Orford et al., 1991; Carter and Orford,

1993; Lorang, 2002). For example, in multiple locations the altitudes of the crests of these depositional features are significantly higher than erosional landforms that are laterally correlated and in close proximity with each other (Fig. 4.3). Currey (1982) is not always clear where the point of measurement for each paleoshoreline was taken; therefore, these uncertainties are unknown.

3) Both the Bonneville and Provo age lakes were threshold controlled; suggesting that geomorphic markers for the SWL of each paleoshoreline should be at a relatively consistent altitude. However, both the Provo and Bonneville lakes have multiple paleoshorelines within a documented range of altitudes. The varying altitudes for these paleoshorelines may be due to high rates of hydro isostatic subsidence during the transgression of the lake (Burr, 1989; Burr and Currey, 1988; 1992). It is difficult to determine which of the various altitudes of the respective paleoshoreline should represent the respective SWL and which paleoshoreline altitude was used at each locality in the dataset.

4) Some geographic areas in the dataset have better spatial distribution than other localities. For example, at both the Provo and Bonneville levels, there are more data points on the eastern side of the basin, and this may bias the interpolation.

5) The Great Basin, and specifically the Bonneville basin, is a tectonically active region, but the amount of tectonically driven adjustment of paleoshoreline altitudes is poorly understood.

Even though the error of individual data points may be significant, the krigged interpolation method used to estimate the water surface averages out the error (Isaaks and Srivastava, 1989). Currey (1982) is the best SWL data currently available, but as we obtain a better understanding of wave processes that can influence the SWL of the lake and obtain more accurate measurements of altitude using differential GPS and LiDAR, the accuracy of the ISM will improve.

The ISM uses an interpolation algorithm to estimate values of hydro isostatic rebound at a given location as a weighted sum of data values at surrounding locations (Isaaks and Srivastava, 1989). Bills et al. (2001) used a spline algorithm to create this interpolated surface; however, a kriging algorithm more accurately and conservatively estimates the rebound in the basin. Spline methodologies tend to overestimate the lows and highs within a database and use a local distance weighted average whereas the kriging methodology tends to produce a smoother surface by using a global weighted average (Isaaks and Srivastava, 1989).

To test for accuracy the ISM was run at a 5 m resolution for both the kriging and spline algorithms for the Bonneville and Provo surfaces. The Bonneville and Provo models were then compared against additional control points collected by the authors. The spline methodology exhibited higher vertical error (~1-2 m) relative to the kriging error (<1 m).

The interpreted altitude of the SWL (i.e., 1,552 masl for Bonneville and 1,444 masl for Provo) is subtracted from each surface to determine the amount of isostatic rebound in the basin. The isostatic correction is then subtracted from the basin's DEM to obtain a new ISM-DEM for both the Bonneville and Provo levels. These ISM-DEMs are time dependent, meaning that the calculated rebound from that surface is only relevant for the time in which the lake resided at a specific altitude. The isostatic correction of the paleoshorelines near the altitude of the Bonneville paleoshoreline will be correctly estimated. However, the isostatic correction will vary with time as the altitude of the still water level and water load change. Contours were extracted from the respective 5 m resolution ISM-DEMs to predict where the lateral extent of the paleoshorelines for the Bonneville and Provo lakes should be located. These contours relate to the interpreted SWL of the respective lake level and act as a visual check for the accuracy of the model by comparing the calculated paleoshorelines with mapped paleoshoreline features evident in aerial photography.

The second correction method uses the equations of Oviatt et al. (1992) to approximate the hydro isostatic rebound of selected profiles in the basin and will be referred to as the Basin Linear Model (BLM). The BLM uses a linear equation to estimate isostatic rebound of paleoshoreline altitudes between the Bonneville and the basin floor (Fig. 4.5). The equation assumes that differential rebound in the basin has a linear proportionality to the water load (depth) at any specific locality.

$$I_z = I_m - (B_m - 1552)\left(\frac{I_m - 1200}{B_m - 1200}\right) \qquad \text{Equation 4.1}$$

I_z is the isostatically corrected altitude of the particular paleoshoreline, I_m is the measured altitude of the particular paleoshoreline, and B_m is the measured altitude of the Bonneville paleoshoreline. The constant 1,552 is the interpreted altitude in m of the Bonneville SWL, and the constant 1,200 is the estimated altitude in m of the basin floor prior to the Lake Bonneville lake cycle. Unlike the ISM model, the isostatic rebound equation is not time specific and estimates the total amount of rebound that has occurred at the given location based on its altitude of the local control points.

Essentially the BLM acts like a broad linear integral to estimate the hydro isostatic rebound at any given point. The model could over or under estimate rebound if the profile at a given locality does not have a linear relationship (Fig. 4.5). Therefore, the BLM was modified in order to focus on the correction of the hydro isostatically corrected altitudes associated with the Intermediate paleoshorelines. This adapted version of the BLM will be referred to as the Intermediate Linear Model (ILM). Because the ILM uses the SWL of the Provo as its lower limit instead of the basin floor, the model better approximates the Intermediate paleoshorelines (Fig. 4.5). The ILM calculates an isostatically corrected altitude.

$$I_z = I_m - (B_m - 1552)\left(1 - \frac{B_m - I_m}{B_m - P_m}\right) + (P_m - 1444)\left(1 - \frac{I_m - P_m}{B_m - P_m}\right) \qquad \text{Equation 4.2}$$

I_z is the isostatically corrected altitude of the particular locality, I_m is the measured altitude of the particular locality, and B_m is the local altitude of the Bonneville paleoshoreline. P_m is the local altitude of the Provo paleoshoreline; the constant 1,552 is the interpreted altitude in m of the Bonneville SWL, and the constant 1,444 is the interpreted altitude in m of the Provo SWL.

Altitude profiles were extracted from representative localities around the basin where Intermediate paleoshorelines were significant (i.e., Hogup / Matlin Mountain region and Wah Wah Valley) using the Bonneville and Provo 5 m resolution ISM-DEMs and an uncorrected 5 m DEM. The altitude profiles from the uncorrected DEM were then corrected using the BLM and ILM. All four of the profiles were compared to each other to determine the difference of the hydro isostatic rebound calculations from the Bonneville and Provo SWLs (Fig. 4.6, Table 4.2).

Hydro Isostatic Model Values and Comparisons

The altitude of the Bonneville and Provo SWLs were measured at multiple localities in the basin (Fig. 4.1, Table 4.1). The measured altitudes of these paleoshorelines trend upward for locations closer to the center of the basin. All the methodologies tested for hydro isostatic correction in this study have limitations and potential for error. The projected paleoshoreline contours from the ISM were calculated with the assumption that they represent the SWL of the lake at its respective level, and that the SWL was essentially a horizontal plane. This assumption is not completely accurate (Atwood, 2006), but is assumed to be a good representation of the average SWL. The lateral variance of the projected location of the paleoshoreline for the 90 n resolution ISM-DEM model was compared to the updated 5 m model. These modeled paleoshoreline locations were compared to paleoshorelines mapped for a geologic map in the Hogup Mountain region (Nelson and Jewell, in review, Chap. 3) and to landforms in aerial photography. For example, on the west side of Hogup Mountain, the lateral variance of the modeled results from this mapped shoreline can vary. The average lateral variance of the modeled vs. mapped paleoshorelines is ~5 m for the 5 m resolution projected paleoshoreline contour and ~40 m for the 90 m resolution model.

The calculated SWL of 1,552 masl for the Bonneville and 1,444 masl for the Provo act as base markers that can be used to determine the variance of each model. The accuracy for each of these models rests on the accuracy of the calculations of the SWLs.

For both the ISM-DEMs, the errors of the isostatic rebound estimates diverge from the calculated surface in question. For example, at the Wah Wah 03 profile, the isostatically corrected altitudes for the ISM-Bonneville and BLM profiles are accurate near the Bonneville SWL but overestimated near the Provo SWL (Fig. 4.6), whereas the isostatically corrected altitude of the ISM-Provo profile is accurate near the Provo SWL but underestimated near the Bonneville SWL.

The BLM uses the local Bonneville altitude and the basin floor as control points; therefore, corrected altitude profiles are accurate near the Bonneville and the basin floor. Depending on the linearity of individual basin profiles, the model could be an appropriate fit for isostatic rebound in the basin. However, if the region does not exhibit a linear fit the model could grossly over or underestimate the rebound (Fig. 4.5). For example, the isostatic correction for the Provo SWL is overestimated by ~2.4 m at the Hogup Spit profiles but is overestimated by ~5.62 m in Wah Wah Valley (Table 4.2). The profile of the basin will never be a true linear fit; therefore, shortening the lateral and vertical distance between the control points will increase accuracy.

The ILM uses the Bonneville and Provo SWLs as control points of the linear equation; therefore, both the local Provo and the Bonneville altitudes need to be identified. Errors from the method result if SWL estimates at these localities are misinterpreted. In comparison to the BLM, the ILM better approximates the isostatic rebound at the Intermediate paleoshorelines by taking a shorter linear fit of the profile. The ILM oversimplifies the effects of hydro isostatic rebound, but as long as the water load is proportional to the isostatic rebound within the region and the profile between the Bonneville and Provo is relatively linear, it does give the best approximation. The ILM estimates the isostatic rebound at the Intermediate paleoshorelines, but the method could also be adapted to estimate isostatic rebound between other lake levels in the basin (e.g., Stansbury or Gilbert levels) by splitting the BLM into smaller integral lengths.

Potential Wave Energy Distribution Model

Projected paleoshoreline extents calculated from the ISM DEMs were used to calculate the fetch of lake surfaces at both the Bonneville and Provo levels. Fetch is often used as a proxy for wave energy; therefore, determining the fetch for these two levels is assumed to approximate the potential wave energy that may be able to reach a specific shorezone. An adapted methodology of Ekebom et al. (2002; 2003) was used to calculate fetch around the basin. Even though fetch is best represented as an area instead of a linear (straight line) relationship, the U.S. Army Corps of Engineers Coastal Engineering Manual (CERC, 2008) suggests that a straight line methodology provides an adequate estimate for fetch and wave parameters. The paleoshoreline extent of each lake's SWL was divided into individual 500 m segments. The straight line methodology calculated the length of the straight line from the individual centroids of each 500 m paleoshoreline segment to another shore projected outward at 5° intervals from 0 to 360°. The maximum and median values from all the fetch directions of each of the centroids were then calculated for the Bonneville and Provo levels. Since specific extents of the Intermediate paleoshorelines around the basin are unknown, the values of the maximum and median fetch for the individual segments at the Bonneville and Provo levels were interpolated together to provide surface rasters for the regions in which the Intermediate paleoshorelines are located. These calculated fetch rasters do not give exact estimations of wave energy that reached a paleoshoreline, but merely aid in estimating the wave energy that could potentially arrive at a specific locality.

The height and type of the landforms vary with the slope of the shorezone because the slope influences the dissipation of the wave energy as the wave moves shoreward. A surface was calculated to estimate the slope of the region in which the Intermediate paleoshorelines are located. This surface was computed in ArcGIS by determining the maximum slope between each of the cells of the 5 m resolution DEM. This surface does not estimate the exact slope present at the time of the lake's occupation in the basin, since it does not account for 1) post-Bonneville erosional or depositional processes that may have changed the slope of the original shore face; 2) changes of slope during the occupation of the lake at that level (the slope is only an approximation of the final shore face surface); and 3) tectonic changes of the slope as the lake level changed.

Suitability models are commonly used in GIS analysis to find favorable locations for a given feature (e.g., locating the best location for a new well, new road or building, or finding the best habitat for bears) by weighting locations relative to given criteria. Therefore, a suitability model within was developed to approximate wave energy potential at a given Intermediate paleoshoreline. In this analysis, the criteria used to determine areas of high or low potential wave energy was a weighted average of rasters expressing the maximum fetch, median fetch, and slope of the regions in which Intermediate paleoshorelines develop. To calculate the weighted average, the rasters were first normalized by reclassifying the criteria into 10 classes related to wave energy potential. According to Adams and Wesnousky (1998), the gradient of relict shore zones control the distribution and type of paleoshoreline features formed. These researchers infer that depositional features such as barrier ridges tend to be deposited in areas of shallow slope (~0–6°), and erosional features such wave cut platforms tend to develop in areas of steeper slopes (> 6°).

The slope raster was reclassified into 10 classes (Table 4.3) based on an adapted 1/3 standard deviation exponential classification scheme. This reclassification scheme fit the best representation of slope ranges suggested by Adams and Wesnousky (1998). Slope is not the only factor than influences landform development; however, Adams and Wesnousky (1998) suggested that barrier ridges tend to be located where the average slope is <4° and wave terraces tend to be located where the average slope is >6°. Each of the fetch rasters was reclassified into 10 classes using the Jenks Natural Breaks algorithm (Table 4.3) which divides the dataset into natural breaks where variation between the values is the greatest. The reclassified rasters were each given a weight (Table 4.3) corresponding to the estimated significance of the specific value to potential wave energy. The weighted rasters were averaged to form a raster surface that estimates the potential wave energy at a specific location.

Potential Wave Energy Distribution

The Bonneville basin has many north south trending normal faults flanking multiple mountain ranges that act as peninsulas or islands that impede the fetch and produce many coves or sub basins that protect the paleoshorelines from high energy waves. Fetch tends to be longer in the main basin of the lake and shorter in the sub basins (Fig. 4.7). The maximum fetch in the basin is 340 km at the Bonneville level and 270 km at the Provo level and are found in the most northerly or southerly portions of the main basin. The highest values of median fetch (97 km) are in the central portions of the main basin and Sevier sub basin (Fig. 4.8).

Wind strength, direction, and duration are very important factors for determining the amount of energy at a specific paleoshoreline; however, the values for these factors during the occupation of Lake Bonneville are currently unknown. With the information currently available, this study's Potential Wave Energy Model (PWEM) is the best substitute for wave energy potential; however it can only act as a first approximation for the conditions at a given site. The kinetic energy within a wave is relatively proportional to the fetch, water depth, wind duration, and wind strength (Atwood, 2006; Nordstrom and Jackson, 2012). If the conditions are optimal, the waves will reach what is called a full developed sea in which the wave energy plateaus and cannot increase. Longer fetches are more likely to have fetch dominated wave trains (where fully developed seas can develop (U.S. Corps of Engineers, 1981; Komar, 1998), whereas shorter fetch lengths are more likely to have fetch limited wave trains (where fully developed seas cannot develop) (Lorang et al., 1993a; 1993b; Atwood, 2006; Nordstrom and Jackson, 2012). Depending on the wind conditions, the main Bonneville basin could produce either fetch dominated or fetch limited wave conditions; however, the sub basins are more likely to produce fetch limited wave conditions.

The distribution of slope in the basin tends to be higher in the central islands and peninsulas of the main basin and on the eastern side of the basin along the Wasatch front (Fig. 4.9). This trend in slope may be due to the active normal faults, such as the Wasatch fault, on the eastern side of the basin (Mohapatra and Johnson, 1998; Velasco et al., 2009), and because the bedrock (i.e., igneous and metamorphic rocks) of the mountains at the eastern end of the basin tend to be resistant to erosion. The shore face slope of the southern sub basins and the northwestern side of the main body tend to have shallower slopes.

Areas within the lake that have a higher predicted energy potential are typically related to regions where erosional landforms are prevalent or where depositional landforms exhibit coarser sediment and higher crest altitudes. Regions with higher wave energy potential also tend to be steeper; therefore, depositional landforms that are present tend to overlie one another, and only the younger or more prominent features are preserved. Areas with lower wave energy potential are typically depositional features that exhibit finer grained material and lower crestal altitudes. Areas of lower wave energy potential also tend to have lower slopes; therefore, the depositional features are laterally spread where more subtle or short lived paleoshorelines are preserved. However, these lower energy areas can also be difficult to study because paleoshorelines near the Provo are often buried or partially buried by fine grained lacustrine deposits as the lake levels rose. Landforms of the Intermediate paleoshorelines seem to be better preserved and more substantial in areas where wave energy potential changed (laterally) over a short distance, had significant sediment supplies, and significant accommodation space. Specific examples of how potential wave energy aids in the correlation of the Intermediate paleoshorelines will be explained in the discussion of the regional case studies.

General Preservation and Development of Paleoshorelines

The wave energy reaching a shoreline is a combination of five basic factors: (1) the fetch, (2) the slope of the shore face, (3) the duration of a storm, (4) the strength of a storm's winds, (5) and the orientation of the storm's winds (Carter and Orford; 1993; Forbes et al., 1995). Using slope and fetch as potential wave energy proxies neglects the role of wind speed, wind orientation, and storm durations. For example, in an area with high slope and high fetch, there is a higher probability that high energy waves reached the paleoshoreline. Without knowing the orientation, the duration, and strength of the regions past winds, there is no guarantee that the predicted wave energy actually occurred. Therefore, these proxies of wave energy potential need to be coupled with other sedimentological and geomorphic proxies to better determine if the region received more energetic waves that could influence the preservation and development of individual paleoshorelines.

The slope of the shore face will influence the dissipation of the wave energy as it moves shoreward. In a shallow shorezone, wave energy is dissipated gradually over a longer distance; therefore, sedimentation increases and you get an even shallower shorezone. In a steep shorezone, wave energy dissipates quickly over a short distance; therefore, there is more wave energy at the shorezone and an increase of erosion. The visual expression of these erosional features and the amount of sediment supply that could be transported from a slope are accentuated by an increasing shorezone bathymetric gradient. This proxy needs to be used with caution because this can be a circular process that as waves erode or deposit sediments the slope of the shore face will also change and as a result causes change in the wave energy.

For this study, fetch was not calculated as a range but by a straight line technique as recommended by the U.S. Army Corp of Engineers. This technique was adequate for the simple approximation needed. However, this method does limit the results because only the median and maximum values are reported and the line calculations can unnecessarily terminate when pathways encounter paleo islands present during both the Bonneville and Provo levels. Depending on the island size, wind strength, and distance from the paleoshoreline centroids, these small islands may or may not have influenced the development of waves and the resultant wave energy. It is unknown what impact these small islands would have on the calculations, so they were included in the calculation. More research on the effect that these islands have on wave development needs to be determined, but is currently out of the scope of this project.

The rate of sediment supply and amount of accommodation space also helps in the development and/or preservation of paleoshorelines. If the sediment supply or accommodation space is low, the features will not be well developed. For example, along the Wasatch Range (east side of the basin), the more resistant igneous and metamorphic bedrock of these mountains causes the sediment supply to be low and the expression of many of these paleoshoreline features are absent or relatively small, except near the mouths of canyons where sediment supply is much greater. In regions where accommodation space and sediment supply is abundant, paleoshoreline features tend to be well preserved. For example, in the Hogup Mountains of northwestern Utah many Intermediate paleoshorelines are preserved due to high sediment supplies from the reworking of alluvial fans and the abundant accommodation space of the mountain range. However, if the sediment rate is too high depositional features may also bury older paleoshorelines; therefore, only the youngest paleoshorelines may be expressed in the profiles of the geomorphic surfaces.

Because wave energy, sediment supply, and accommodation space all differ from region to region, many paleoshorelines may not be preserved or may not ever develop at some localities. The complexity of the paleoshoreline development is directly proportional to the length of time or duration in which the SWL was near the altitude of the individual paleoshoreline. However, in areas where accommodation space, wave energy, and sediment supply are significant, simple depositional features can develop very quickly. For example, 3–4 m high barrier ridges in Pyramid Lake in western Nevada formed in less than six months (Adams and Wesnousky, 1998), and strandlines formed during the brief (6 month) existence of Lake Thistle, Utah in 1983 (Costa and Schuster, 1988). Storm events tend to bring in most of the wave energy at these features and do most of the erosion, deposition, or transport of sediments (Anthony, 2006; Weir, 2006). Even though depositional landforms can develop quickly, they do not develop uniformly due to differing topographical factors and the heterogeneous nature of wave energy. Most of the sediments of the Intermediate paleoshorelines tend to be composed of coarse material due to the short time in which these sediments were continually eroded at a constant altitude, the winnowing of finer sediments offshore, and the relative proximity of the sediment sources.

Intermediate Paleoshoreline Correlations

Since the basin is so large, the correlation of the Intermediate paleoshorelines focused on two (2) regions (i.e., Hogup/Matlin Mountains and Wah Wah Valley) where

Intermediate paleoshoreline features are prominent and abundant. Figures 4.10 – 4.13 show the altitudinal profiles and aerial photography for these areas. The regions were selected because of their distal spatial distribution, multiple profile locations where local correlations can be investigated, and disparate conditions of isostatic rebound and wave energy environments. To determine the SWL of the features, the isostatically corrected altitudes of erosional notches and the inflection points and/or maximum crest heights were measured and compared to each individual paleoshoreline in the profile. Six prominent Intermediate paleoshorelines are seen throughout the study areas. The numbering scheme for these features is based on the best represented profile in the Matlin embayment of the Matlin Mountains (Fig. 4.10) and is similar to numbering schemes for other paleoshorelines in the basin (Burr and Currey, 1988; Godsey et al., 2005). Figure 4.14 shows the maximum crest heights of the depositional crests or the inflection point of the erosional platforms related to the Intermediate paleoshorelines. The variance of the altitude for each of the paleoshoreline crests and/or erosional notches from each profile is reported in Fig. 4.14 along with the median and standard deviation of the altitudes compared to the average measured SWL of each paleoshoreline.

Hogup and Matlin Mountains

The Hogup/Matlin Mountains are in the northwestern portion of the Bonneville basin bordering the northwestern shore of the Great Salt Lake. The region has long fetch and a topography of steeper bedrock headlands with multiple small embayments. Gilbert (1890) described multiple Intermediate “cut and built” terraces (Russell, 1885) on the south slope of Matlin Mountain that Gilbert referred to as the “Dove Creek

Embankments." Besides these depositional terraces at Dove Creek, other well preserved Intermediate paleoshoreline features (e.g., wave cut platforms, barrier ridges, baymouth bars, tombolos, and strandlines) in the region provide a rich diversity of depositional and erosional settings. The region had abundant sediment supplies from large pre Bonneville alluvial fans and highly fractured local bedrock (dominated by the quartzites, limestones, and cherts of the Pennsylvanian and Permian Oquirrh Group). Steep shore zones associated with the bedrock headlands provided excellent locations for erosional landforms (e.g., wave cut platforms) to develop. The sediment from these erosional landforms was transported to protected embayments where substantial depositional landforms accumulated due to the abundant accommodation space and change in wave energy.

Matlin/Dove Creek Profiles

In the Matlin embayment (Fig. 4.10), six distinct Intermediate barrier ridges are preserved. Based on the geomorphic shape of the barriers, the prominent longshore transport direction of sediments is to the northeast. The PWEM suggests that the paleoshorelines west of the embayment have high potential wave energy due to their relatively steep slopes and long fetch (Fig. 4.15). The topography of the headlands on both sides of the embayments focused the waves toward these headlands, and the change in wave energy caused the sediments to be deposited in the shallow slope of the embayment. The altitude of the six prominent Intermediate ridges in the embayment are consistent with the altitude of similar isostatically corrected paleoshorelines around the basin (Fig. 4.14). Other paleoshorelines are recorded in the sediments of the embayment;

however, there is no visual surface expression of the features. The higher sediment supply and abundant accommodation space in the embayment caused younger paleoshorelines to bury older paleoshorelines. The young paleoshorelines are more prominent because they inherited the shape and size of former paleoshorelines. These more prominent Intermediate paleoshorelines are numbered from lowest to highest in altitude I-1 through I-6. This numbering scheme follows the numbering schemes of other paleoshorelines within the basin. The exact chronologic order of the Intermediate paleoshorelines are unknown, but it is assumed that the lower paleoshorelines are older. The altitudes of the barrier crests of paleoshorelines I-1 to I-6 tend to be ~1–2 m higher in the Matlin embayment than on barrier crests in the Dove Creek profiles located on the other side of Matlin Mountain (Fig. 4.14). This confirms the patterns of the PWEM and suggests that higher wave energy originated from the south, possibly due to the longer fetch for the Matlin paleoshorelines. I-1 at Matlin is partially eroded by waves of the younger Provo age lake.

The Dove Creek profiles were measured from a series of six to seven prominent depositional barriers and spits on the north side of Matlin Mountain (Fig. 4.10). Like the Matlin embayment, the PWEM and the geomorphic expression of the features (i.e., orientation of spits and bars) suggest a prominent northeasterly direction of longshore sediment; however, direction of longshore drift changed to a northwesterly direction on the north side of the mountain. Bonneville age erosional wave platforms on the southeast slope of Matlin Mountain are ~3–4 m lower than the corresponding depositional crests suggesting that the storm run up in the region was significant (Fig. 4.3b). In addition to the six Intermediate paleoshoreline markers seen in the Matlin embayment, some smaller paleoshorelines are identified at Dove Creek. In the Dove Creek 01 profile, two small paleoshorelines (one between I-4 and I-5 and one between I-3 and I-4) are distinctly seen, but are not identified in the Matlin profile. Faint strandlines on the erosional paleoshorelines west of Matlin embayment suggest that these features may exist near the Matlin profile; however, the paleoshorelines may have been buried by the bigger I-4 and I-5 depositional features. At the Dove Creek 02 profile, two additional paleoshorelines markers are observed (one between the I-2 and I-3 paleoshorelines and another one directly above I-4 – Fig. 4.14).

Big Pass Profiles

During the transgression of the lake, sediments from alluvial fans south of the Big Pass embayment were excavated by wave action and transported northward via longshore drift (Fig. 4.11 and 4.16). The embayment of Big Pass is relatively well protected by its flanking bedrock headlands allowing the formation of multiple Intermediate gravel barrier ridges. These barrier ridges exhibit progradational sedimentation patterns, interpreted to be due to the high rate of sediment supply, which outpaced the transgressive rise (rate of accommodation space) of the lake (Fig. 4.16). The gravels near the crests of the beach ridges are often cemented with calcium carbonate cement (beachrock) in shallow foresets that steepen down the slope of the ridge. Lagoonal silts and sands are typically draped behind these ridges followed by thin layers of deepwater marl. Very thin gravel lenses, interpreted as wash over gravels, can be found within these marls and lagoon deposits. As the lake continued to rise, transgressive marls overlaid the gravel barrier ridges; however, most of the marls eroded away following the forced regression of the Bonneville Flood. The same six Matlin paleoshoreline (I-1 through I-6) markers are seen at this locality; however, at a slightly lower altitude than at Matlin (Fig. 4.14). It is assumed that the gentler slope and shorter fetch for these paleoshorelines caused less wave run up in the embayment. Paleoshorelines I-1 to I-5 tend to be more significant than I-6. Based on the topography and the sediment sources at the relative altitudes of the paleoshorelines, it is assumed that the lower features had a higher rate of sediment supply and more accommodation space available than the upper features.

In most localities of the Hogup/Matlin Mountain region and specifically at Big Pass embayment, paleoshoreline I-1 exhibits a large beach ridge that has been incised by the regressive Provo paleoshoreline (Fig. 4.17). Gastropod samples collected from the I-1 sediments within the embayment date at 18,990 ± 190 ^{14}C yr B.P. (Sack, 1999). Transgressive features described in quarries near North Salt Lake, Utah (Scott, 1988) have the same ages, relatively similar isostatically corrected paleoshoreline features, and similar stratigraphic relationships to the Provo level as seen at Big Pass. Therefore, it is assumed that the Big Pass and North Salt Lake features are both I-1 paleoshorelines and provide some radiometric control for the lowest Intermediate paleoshoreline set. The age of these paleoshorelines relate to the timing and altitude of the U1 oscillation of Oviatt (1997). Much like the topography of the Matlin embayment, the headlands on both sides of the embayment were the focus of higher energy waves causing a change in wave energy that induced sediments to be deposited on the shallow slope of the embayment. All of the same less prominent Intermediate paleoshorelines seen in the Dove Creek profiles are seen in the Big Pass region except one (Fig. 4.14). However, one additional paleoshoreline can be seen in both of the Big Pass profiles in between I-5 and I-6.

Wah Wah Valley

Gilbert (1890) discusses multiple substantial Intermediate spit complexes in the Wah Wah Valley that are oriented to the southwest (Fig. 4.18). Each spit complex has up to eight distinct spits and terraces at various elevations. These spit complexes are located on the southerly capes of alluvial fans on the flanks of the Frisco Mountains that provided an excellent sediment source as the lake rose and reworked the alluvial deposits. Between the altitudes of the spit crests, multiple strandlines can be delineated marking various tracks of southerly longshore drift. The accommodation space was low; therefore, significant paleoshorelines do not develop. The Wah Wah Valley has a low relative fetch potential. The median fetch values range from 8–11 km and max fetch ranges from 113–135 km. Slopes in the region are relatively low with an average slope of < 3°. However, the slopes on the western side of the valley are as high as 6°. The hydro isostatic rebound in the region is also low due to the relatively shallow depth of the lake.

Four different altitudinal location profiles were measured in this region (Fig. 4.18), two from the west (Fig. 4.12) and two on the east side of the valley (Fig. 4.13). In the Wah Wah Valley, all of the significant paleoshorelines seen in the Hogup/Matlin Mountain region (I-1 through I-6) can be readily seen except I-1. I-1 was either buried by finer grained marls or never strongly developed during the transgressive phase of the lake. The PWEM suggests that the region had very low wave energy potential at lower altitudes due to the gentle slope and short medium fetch. I-2 is seen in the region but difficult to detect, since the features tend to be partially buried or smaller than its northern equivalents. Sediment supply, accommodation space, and wave energy potential all increased within the valley as the SWL rose to the Bonneville level. The upper paleoshorelines I-3 to I-6 are progressively larger than the lower paleoshorelines since the sediment load and wave energy potential increased with the water level rise. The profile localities in the eastern side of the valley had such a high sedimentation rate and relatively steep slope (~4–6°) that many individual paleoshorelines overlap each other making it difficult to determine distinct paleoshorelines. The profile localities on the west side of the valley have a relatively moderate slope (~1–4°) where the individual paleoshorelines spread out laterally and were better preserved. The shallow nature of the sub basin and the high rate of sediment supply may have allowed many of the less substantial paleoshorelines to be well preserved.

Wah Wah West Group

The Wah Wah 01 and 02 profiles are on the western side of Wah Wah Valley (Fig. 4.18). Wah Wah 01 is in the northern part of the valley in a headland protected embayment (Fig. 4.12a) and Wah Wah 02 is in the southwestern portion of the basin (Fig. 4.12b) where fetch was the highest in the valley. At the Wah Wah 01 locality the PWEM (Fig. 4.18) suggests very low wave energy potential, although multiple depositional paleoshorelines are well preserved. The prominent depositional features are preserved due to strong longshore currents, an abundant sediment supply, and high accommodation space available in this small embayment. I-1 is not detected in the profile, I-2 is only faintly preserved, but I-3 through I-6 are prominently preserved. In addition to the prominent Matlin paleoshoreline markers, multiple other paleoshorelines can be seen in the profile. The two (2) smaller paleoshorelines between the 1-4 and 1-5, the paleoshoreline between I-6, and the Bonneville s present; however, another paleoshoreline is also noted between I-5 and I-6 (Fig. 4.14). Other paleoshoreline features are seen in the profile but they cannot be correlated to other sites in the region. The individual paleoshorelines are more laterally spread out and better preserved in the record because of their very shallow slope (< 2°) and relatively coarse sediment (i.e., McMillan and Teller, 2012; Tamura; 2012). As the lake rose, the shoreline transgressed over the long lateral distance and the wave base could not readily erode or build on lower paleoshoreline features.

The Wah Wah 02 locality is predicted to have had a higher wave energy potential due to the relatively steep slopes (2–4°) in the region and due to the openness of the region to the Sevier basin (longer fetch). As predicted by the PWEM and the geomorphic orientation of the paleoshorelines spits and beach ridges, longshore sediment transport moved the sediment in a southerly direction (Fig. 4.18). The profile locality is on the edge of a bedrock headland where wave energy would be reduced as the water depth increased. The reduced wave energy, high accommodation space and relatively high sediment supply allowed substantial Intermediate features to form. I-2 through I-6 paleoshorelines are clearly seen in the profile; however, only one smaller paleoshoreline between I-4 and I-5 is preserved (Fig. 4.14).

<u>Wah Wah East Group</u>

The two profiles of the east group have higher fetches (92–113 km) and slopes (~4–6°) than the western profiles. The longshore transport direction is to the southwest and sediment was derived from large pre-Bonneville fans (Fig. 4.13 and 4.18). The individual Intermediate paleoshorelines form a series of spit complexes on the southern portion of the capes of alluvial fans. Due to the high rate of sediment supply, many of the features in this group tend to be partially buried by other paleoshorelines; therefore, it is difficult to clearly distinguish between some individual paleoshorelines, which may account for variations in the crestal heights at these localities. I-3 through I-5 are clear in both the Wah Wah 03 and 04 profiles (Fig. 4.17). I-2 is only faintly observed in both of the profiles because it is partially buried by finer grained deep water sediments. I-6 cannot be clearly distinguished in the Wah Wah 03 profile. The higher slope (~6°) and abundant sediment supply near the Bonneville level probably caused sediments related to the Bonneville paleoshoreline to bury remnants of I-6. In the Wah Wah 03 profile there are two of the smaller paleoshorelines in between I-4 and I-5 paleoshorelines and another paleoshoreline between I-5 and I-6 (Fig. 4.14). These same smaller paleoshorelines are seen in the Wah Wah 04 profile except for the lower paleoshoreline in between I-4 and I-5. Some of the lower paleoshorelines in the region appear more dissected by ephemeral gullies and have evidence of marl deposits interfingered between overlying gravel deposits. Gilbert (1890) suggested that these lower paleoshorelines could be evidence of an older lake cycle; however, these deposits may also be evidence of an oscillating lake surface during the transgressive rise of the lake. Without a more detailed stratigraphic and radiometric analysis this chronology cannot be established.

Conclusions

All three of Gilbert's hypotheses (hydro isostatic rebound, differential wave energy, and differential autogenic processes of an oscillating lake level) play a role in the altitudinal variations of the Intermediate paleoshorelines. Hydro isostatic correction of the paleoshoreline altitudes is vital for understanding the broad correlations of the features; however, it does not account for smaller variations in altitude or the preservation of paleoshorelines in the geomorphic record. Of all the methodologies tested for the correction of hydro isostatic rebound, this study's ILM generates the most accurate corrections of the Intermediate paleoshorelines. The ILM probably over simplifies the hydro isostatic rebound, but by using this first approximation, multiple paleoshorelines can be correlated in the basin. The ILM model is dependent on the accuracy of the SWL measurement for both the Bonneville and Provo levels. The SWL measurement can be very difficult to assess and will be the main source of error in the ILM. An adapted ILM model could be very useful in better approximating isostatic rebound (by using shorter an integral distance) in other paleoshoreline intervals of the Bonneville basin (e.g., between the Stansbury and/or Gilbert). However, the post Bonneville tectonic history of faults in the basin also need to be better assessed in order to understand how the hydro isostatic rebound models need to be adjusted.

The GIS techniques of the PWEM model improves understanding of the potential wave energy at each site. The values of fetch and slope used in the PWEM was successfully used as a proxy for potential wave energy and can help predict and explain the existence of erosional or depositional landforms, give potential reasons for the variation of the altitudinal extents of a paleoshoreline, and aid in the explanation of the preservation or erosion of specific paleoshorelines. The PWEM only suggests the potential wave energy conditions at a paleoshoreline and the lack of data on storm patterns limits the full usefulness of the proxy. However, the autogenic sedimentological processes of how the deposits of an oscillating lake are preserved differently in one region compared to the next can be better understood if the PWEM model is used in conjunction with a stratigraphic and geomorphic understanding of a local region.

Based on the comparison of the altitudinal limits and sedimentological evidence seen for the Intermediate paleoshorelines within these field sites, it appears that better hydro isostatic rebound calculations, accounting for variations due to wave energy, sedimentological, and geomorphic factors have make it possible to correlate the features in the basin. At the test case localities, six significant paleoshorelines have been correlated locally and basin wide (Fig. 4.12). Smaller scale paleoshorelines are also seen in the record and can be correlated locally or basin wide, depending on the sedimentological and geomorphic conditions at the locality. Variations in the altitude of individual paleoshorelines can be accounted for by looking at the PWEM coupled with other known sedimentological and geomorphic proxies.

With the correlation of the Intermediate paleoshorelines now better established, the next step would be to look at the sedimentological evidence and dating of individual paleoshorelines to determine their chronology. There are multiple radiometric age dates for paleoshorelines near the main levels of the lake (i.e., Stansbury, Bonneville, Provo, and Gilbert), but there are not many radiometric ages, with good stratigraphic contexts, that record the Intermediate paleoshorelines (Fig. 4.19). Oviatt (1997) suggests that three (3) large oscillations (U1 – U3) may have occurred during the transgression of the lake from the Stansbury to the Bonneville level. The U1 oscillation (Fig. 4.19) is within the upper altitudinal range of I-1, the U2 oscillation is in the attitudinal ranges of I-3 through I-5, and U3 oscillation is the altitudinal range of I-5 through I-6. In addition, the sedimentological and geomorphic evidence in the Hogup Mountains and in Wah Wah

Valley suggests that two additional large oscillations may have occurred during the I-4 though I-6 paleoshorelines (Nelson and Jewell, in review, Chap. 5). Even though there are many indications that the intermediate paleoshorelines do record sub millennial changes in climate, until more information is obtained it will remain unknown how specific paleoshorelines relate to each of the oscillations.

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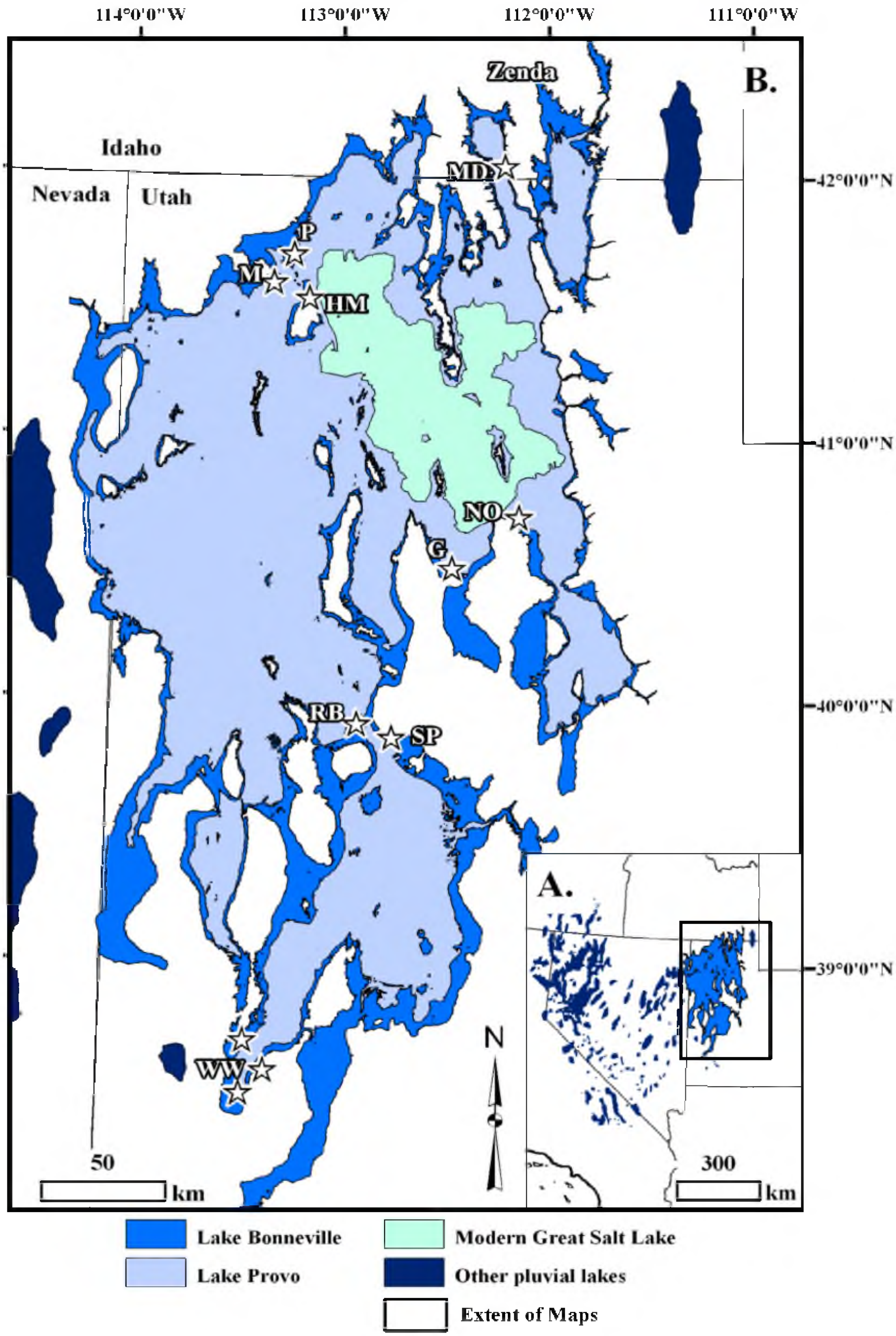
Figure 4.1: Extent of the modern Great Salt Lake in relation to the extent of Pleistocene Lake Bonneville and Lake Provo; and the relation of where the field localities are within the basin. Matlin Mountains (M), Hogup Mountains (HM), Peplin (P), Malad Pass (MD), North Oquirrh (NO),Grantsville Spits (G), the Snowplow (S), Reservoir Butte (RB), and Wah Wah Valley (WW).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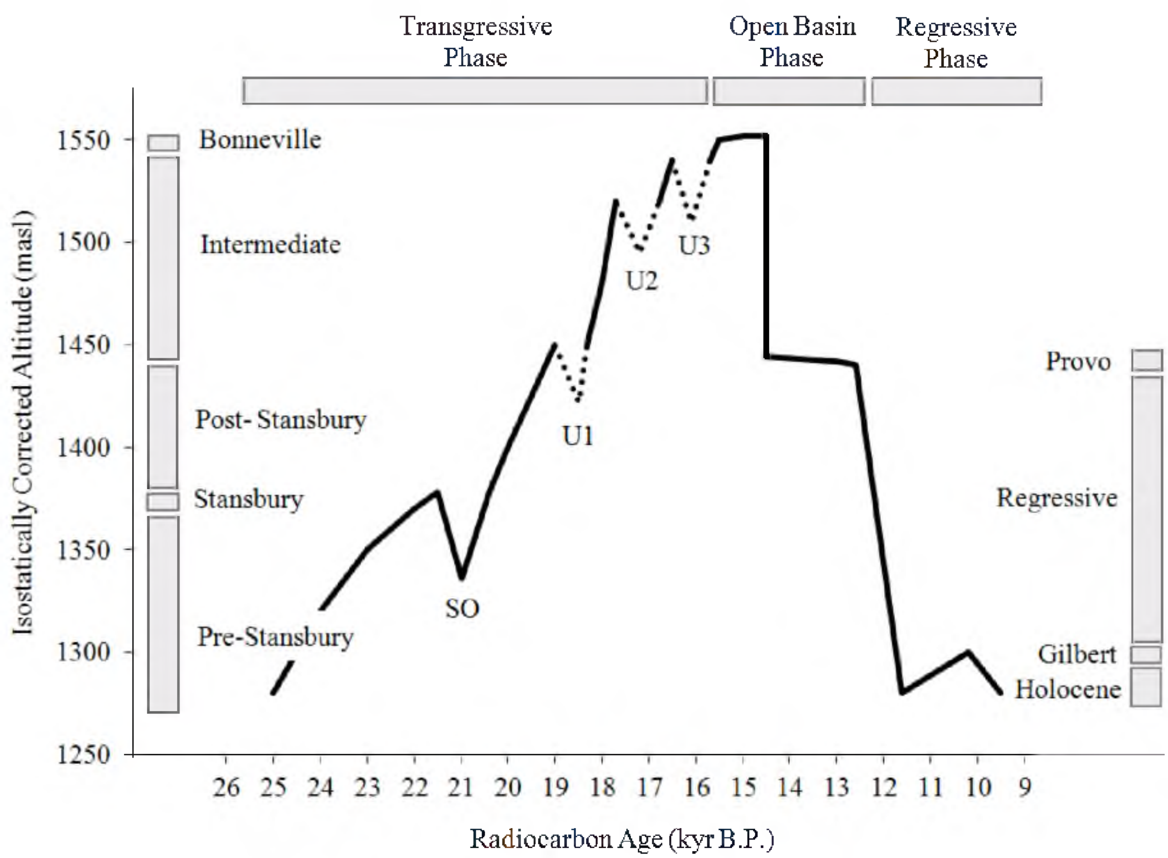


Figure 4.2: Lake Bonneville hydrograph modified from Oviatt (1997) and Godsey et al. (2011). Altitudes are adjusted for effects of differential isostatic rebound in the basin (Oviatt et al., 1992). Amplitude limits of lake-stage fluctuations associated with the U1, U2, and U3 oscillations are approximate and are shown here schematically. The temporal range of the transgressive, regressive, and open phases of the lake cycle are shown horizontally, whereas the altitudinal range of each of the paleoshorelines is shown vertically within either the transgressive or the regressive phases of the lake cycle.

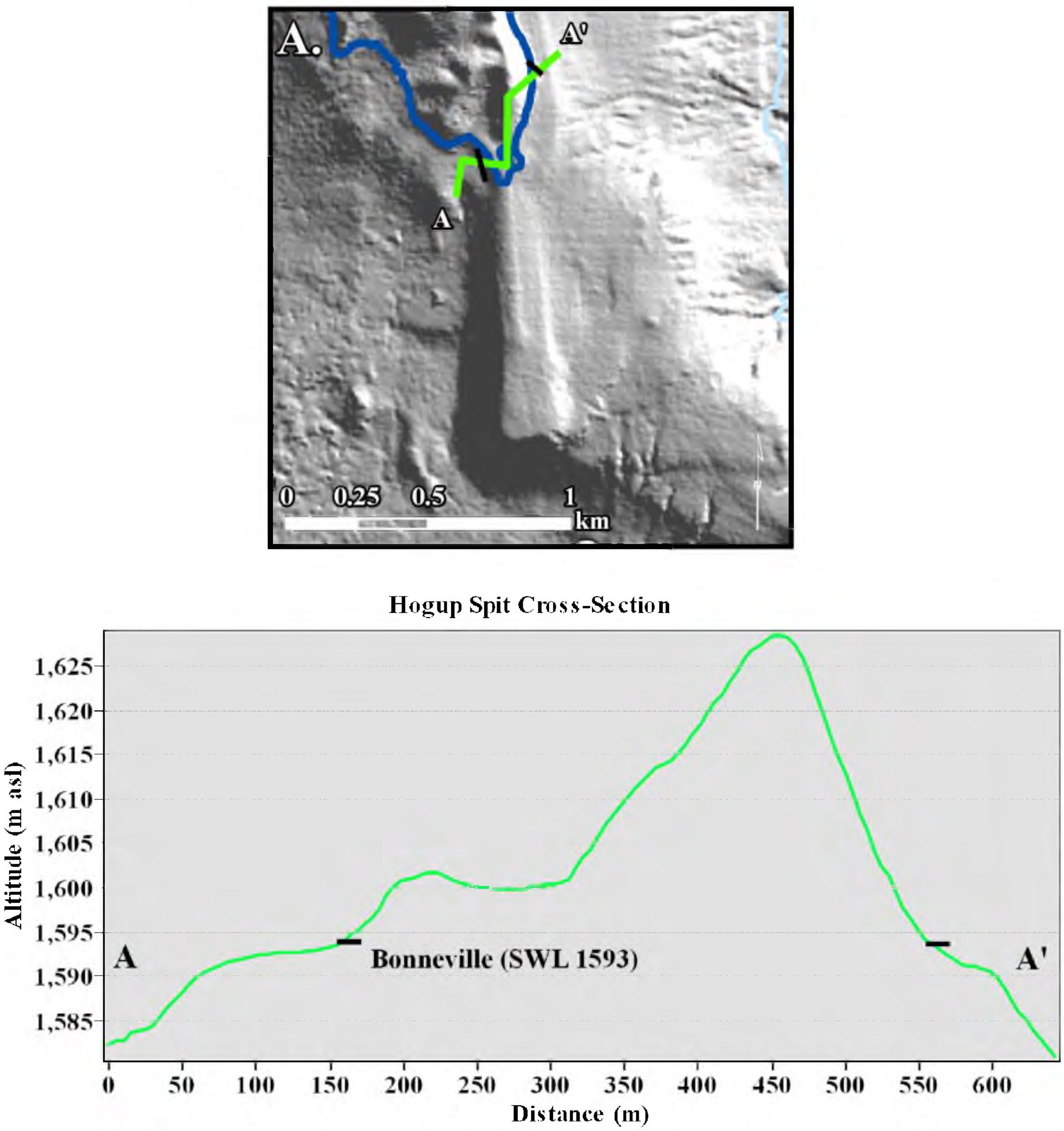


Figure 4.3: Cross section profiles of selected localities that exhibit elevated depositional landforms above the SWL estimated at the inflection point of an erosional wave platform. A: Hogup Spit; B: Dove Creek; C: Reservoir Butte. Black markers designate the SWL.

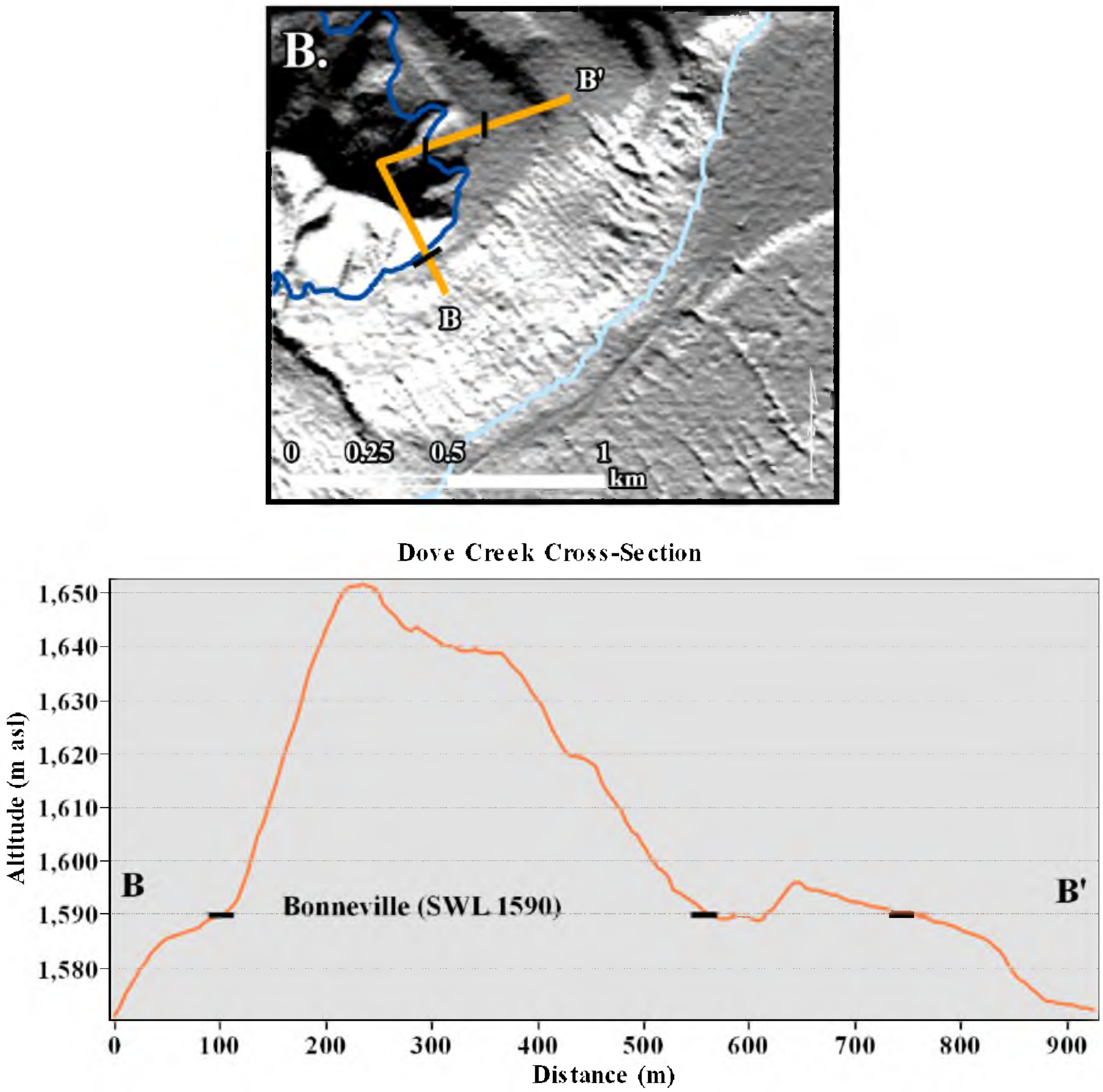
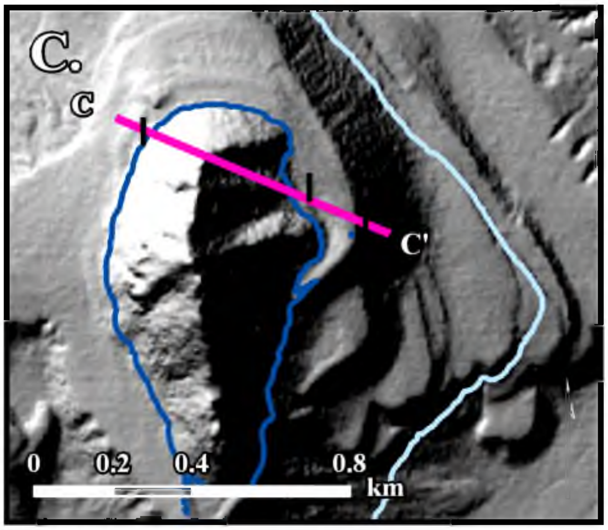


Figure 4.3 (cont.)



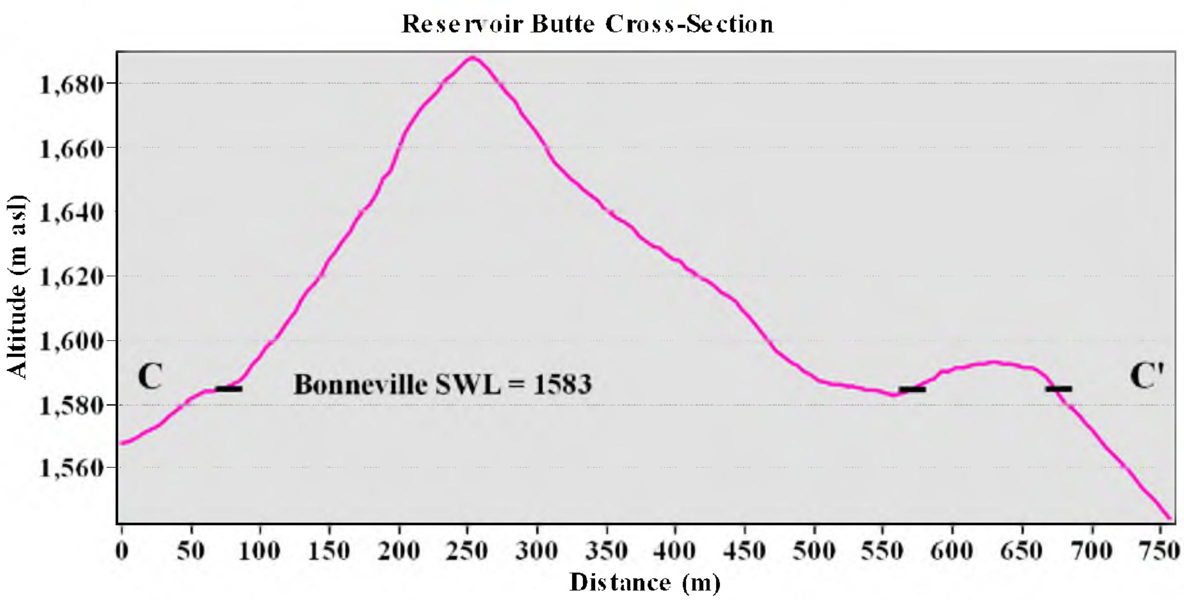


Figure 4.3 (cont.)

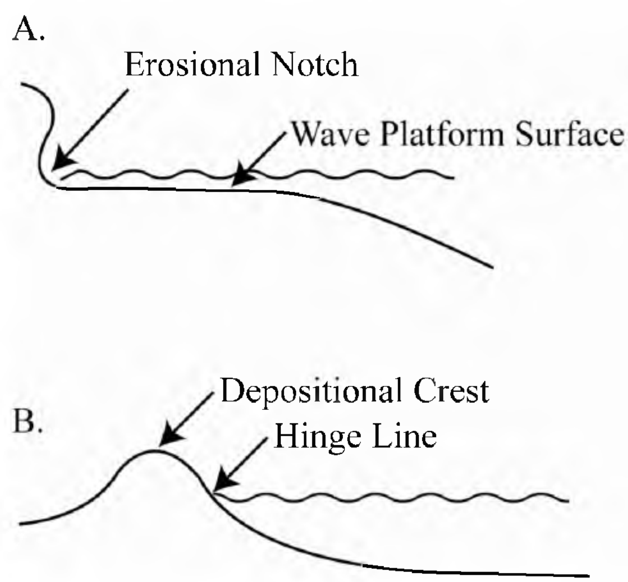


Figure 4.4: Schematic locations of SWL estimates.

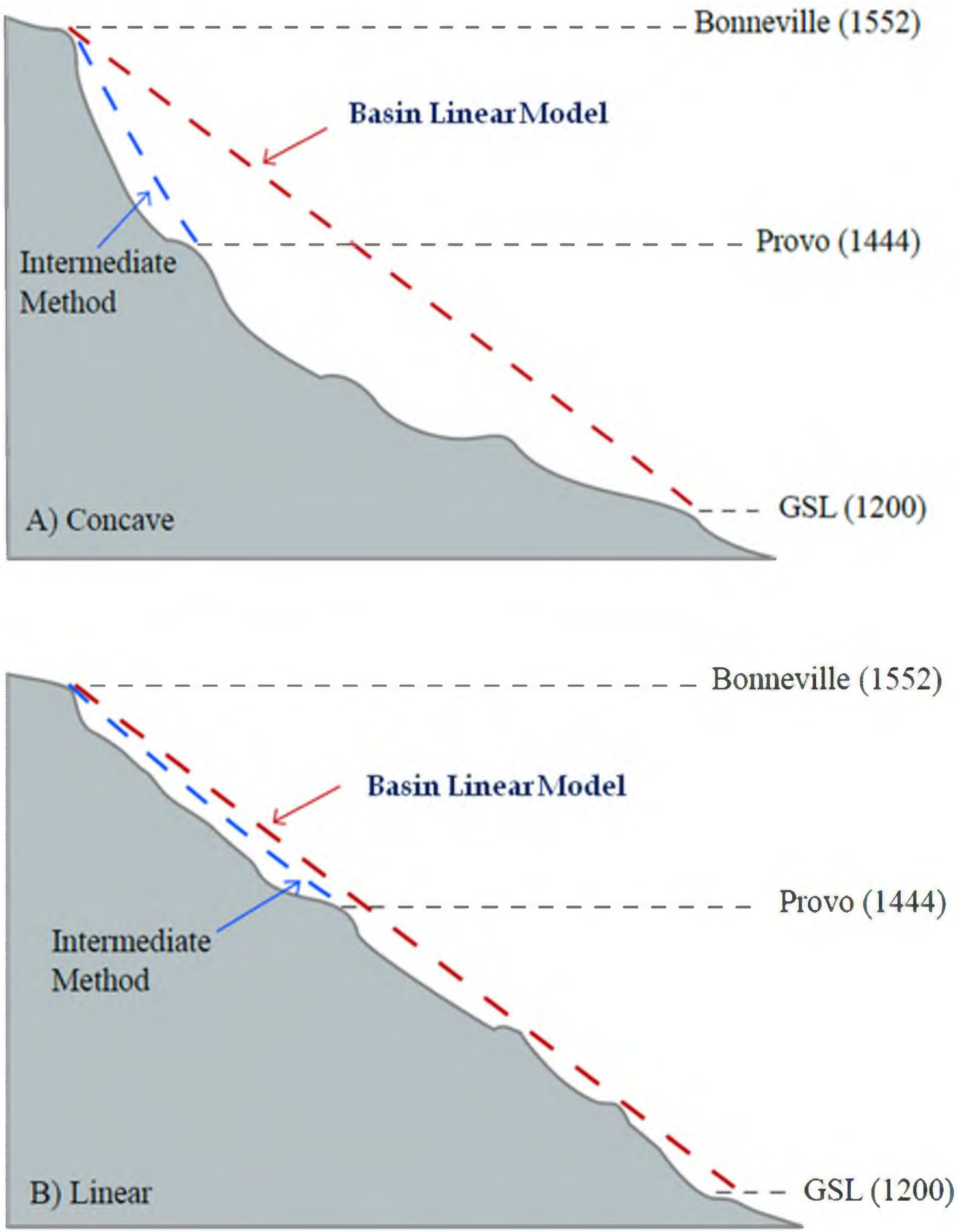
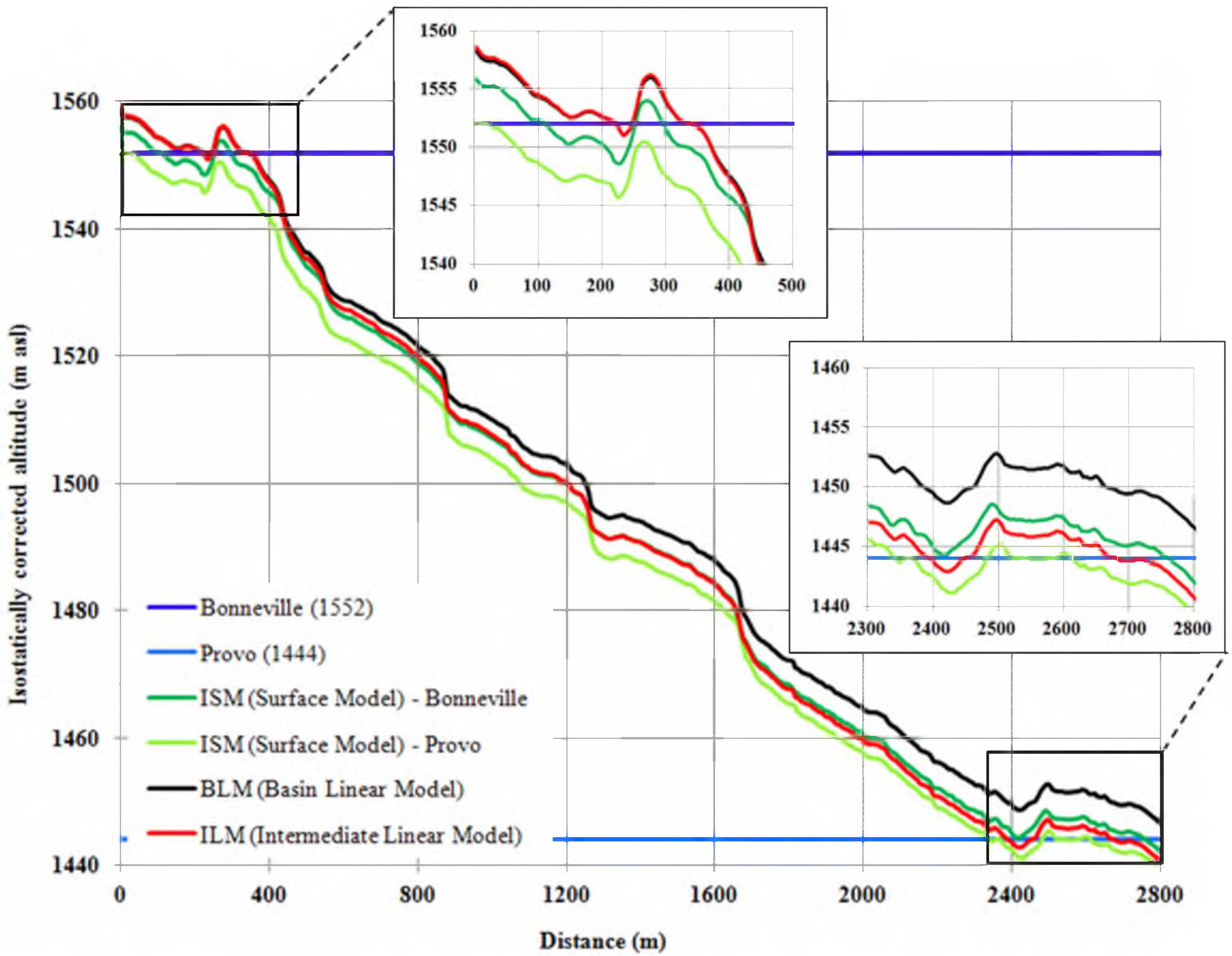


Figure 4.5: Schematic drawing of implications of the Basin Liner Model in comparison to the Intermediate Linear Model.

Figure 4.6: Visual divergence of the various hydro-isostatic rebound correction models from the Bonneville and Provo SWL's based on the Wah Wah 04 altitude profile.


Isostatically corrected altitude (m asl)
Distance (m)
1560
1540
1520
1500
1480
1460
1440
0
400
800
1200
1600
2000
2400
2800
Bonneville (1552)
Provo (1444)
ISM (Surface Model) - Bonneville
ISM (Surface Model) - Provo
BLM (Basin Linear Model)
ILM (Intermediate Linear Model)
1560
1555
1550
1545
1540
0
100
200
300
400
500
1460
1455
1450
1445
1440
2300
2400
2500
2600
2700
2800

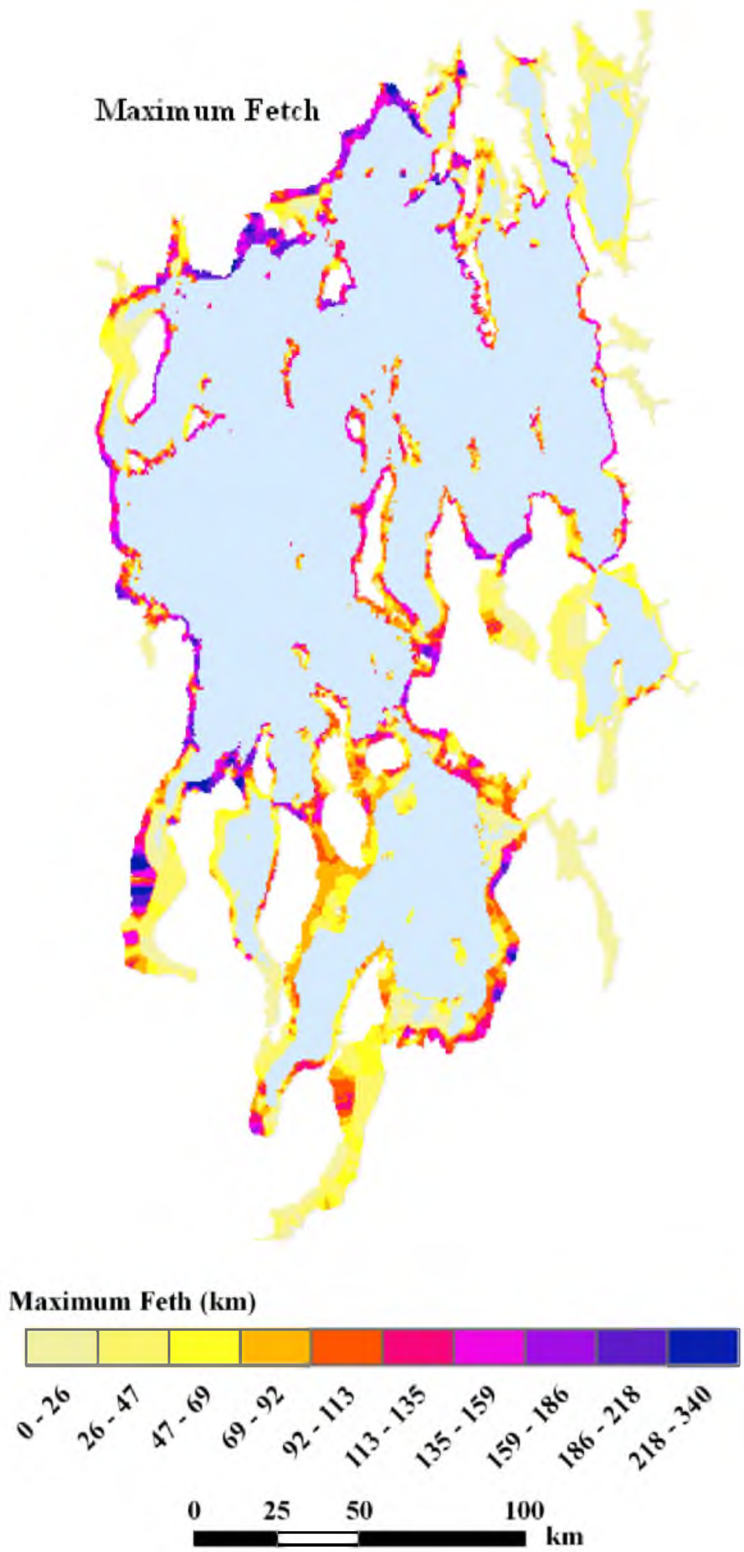


Figure 4.7: Maximum fetch distribution for the Intermediate paleoshorelines.

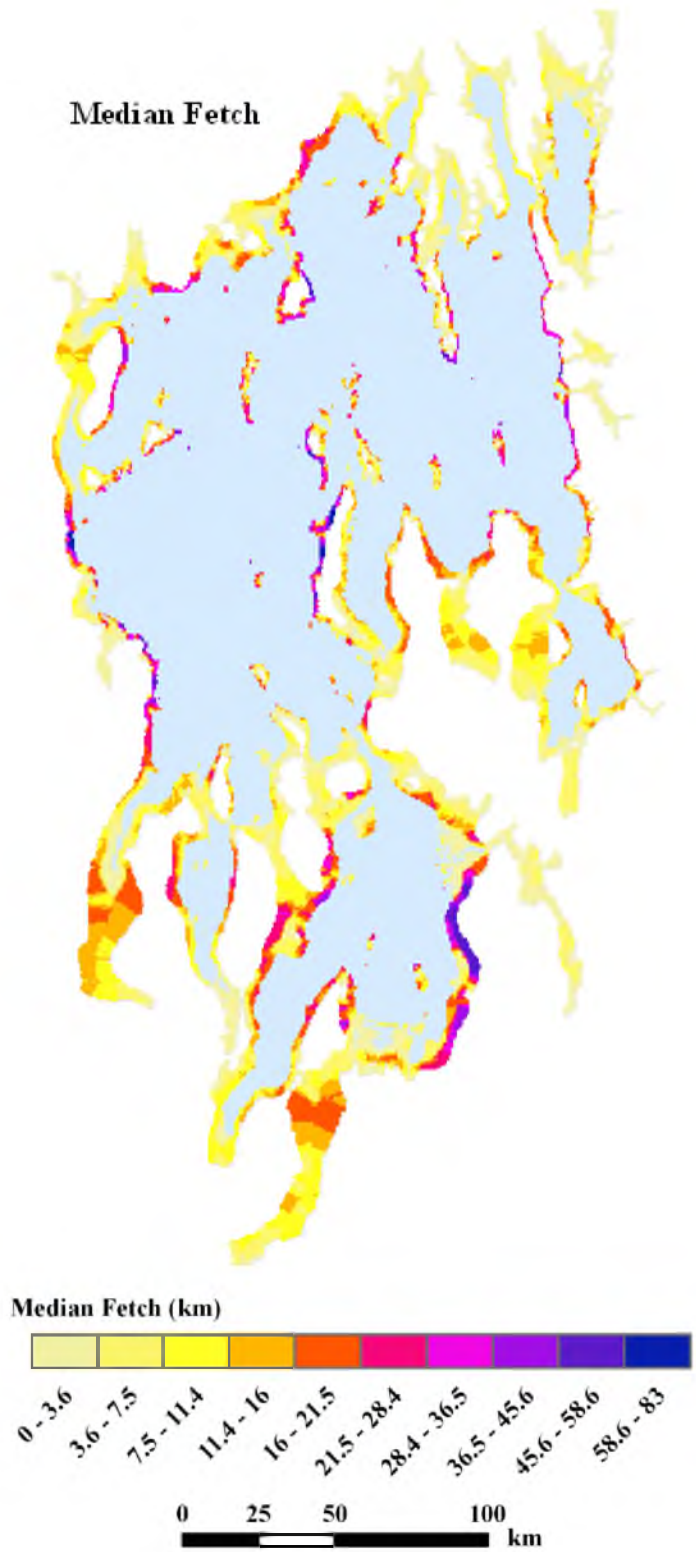


Figure 4.8: Median fetch distribution for the Intermediate paleoshorelines.

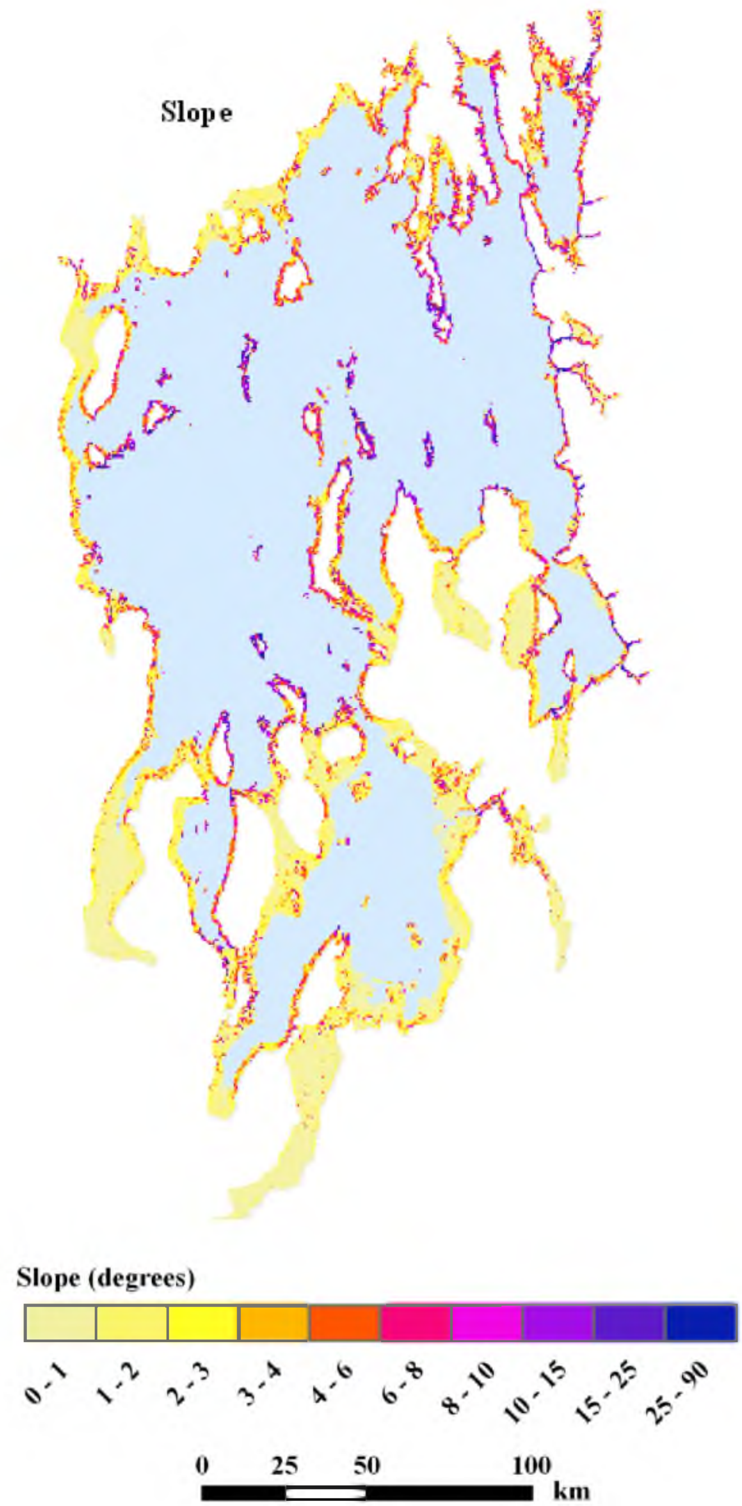


Figure 4.9: Slope distribution for the Intermediate paleoshorelines.

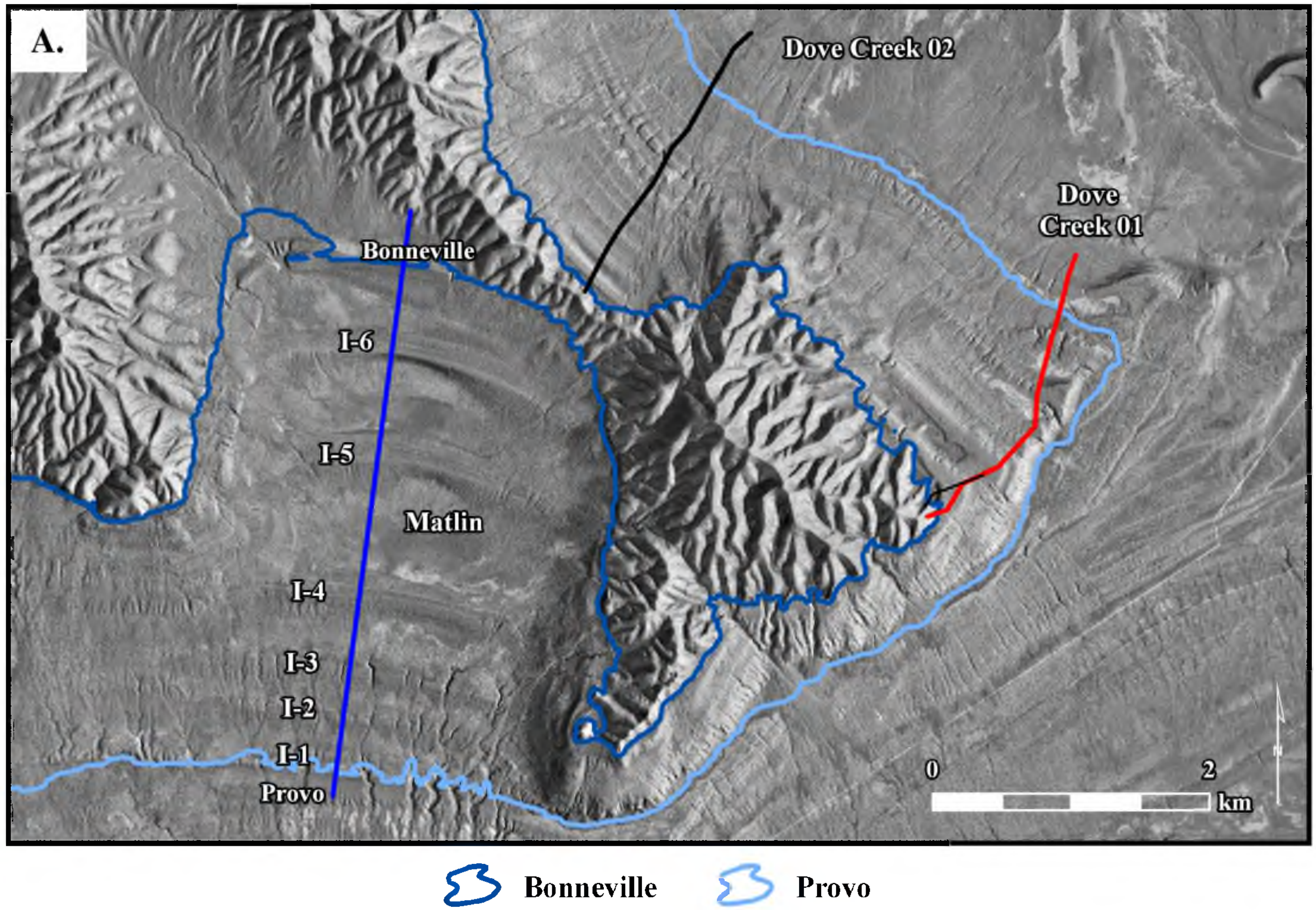


Figure 4.10: Matlin Mountain and Dove Creek altitudinal profiles corrected with the ILM. Profiles start slightly above the Bonneville level and end below the Provo level. A) Aerial photography and profile locations, B) Attitudinal profiles.

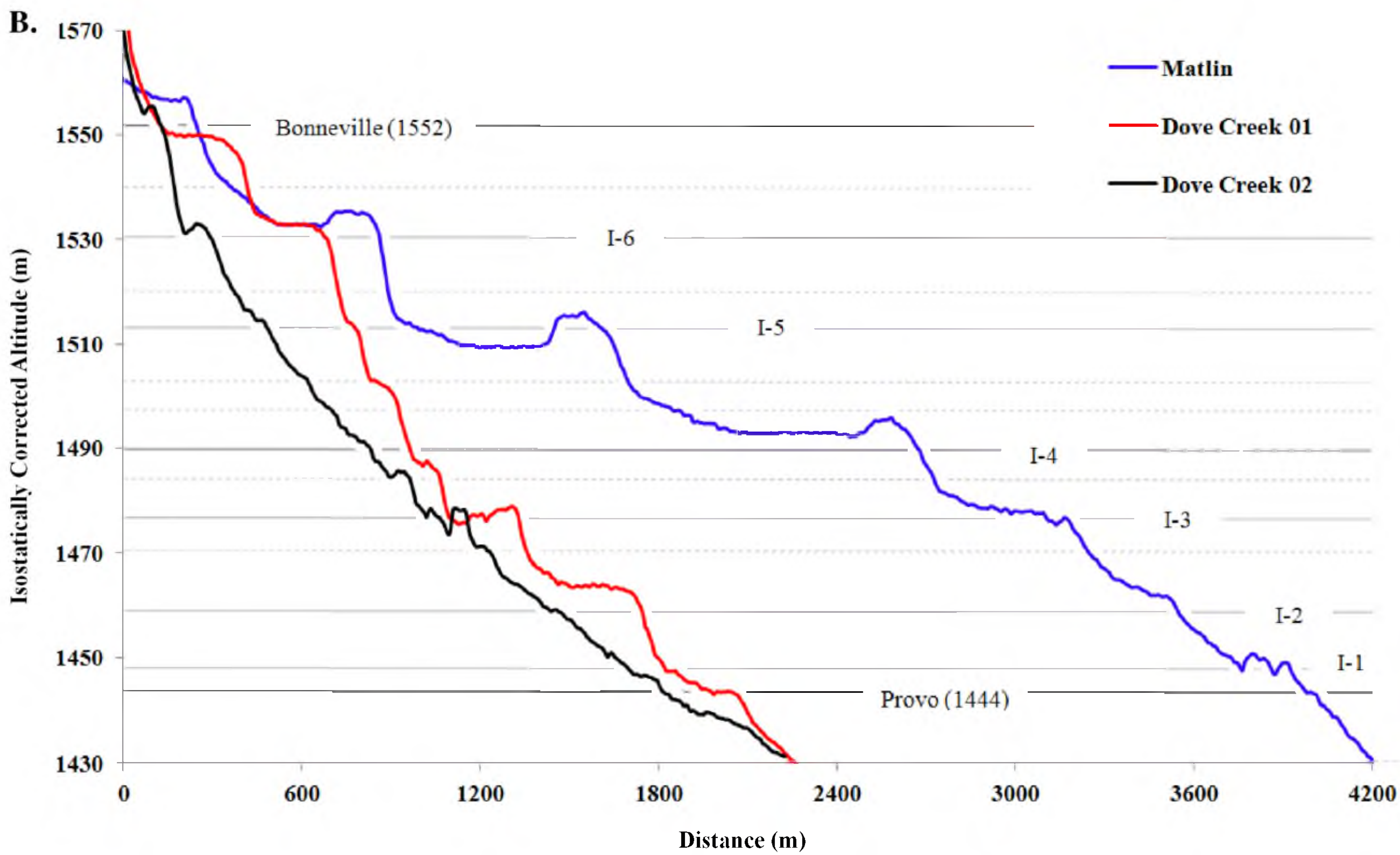


Figure 4.10 (cont.)

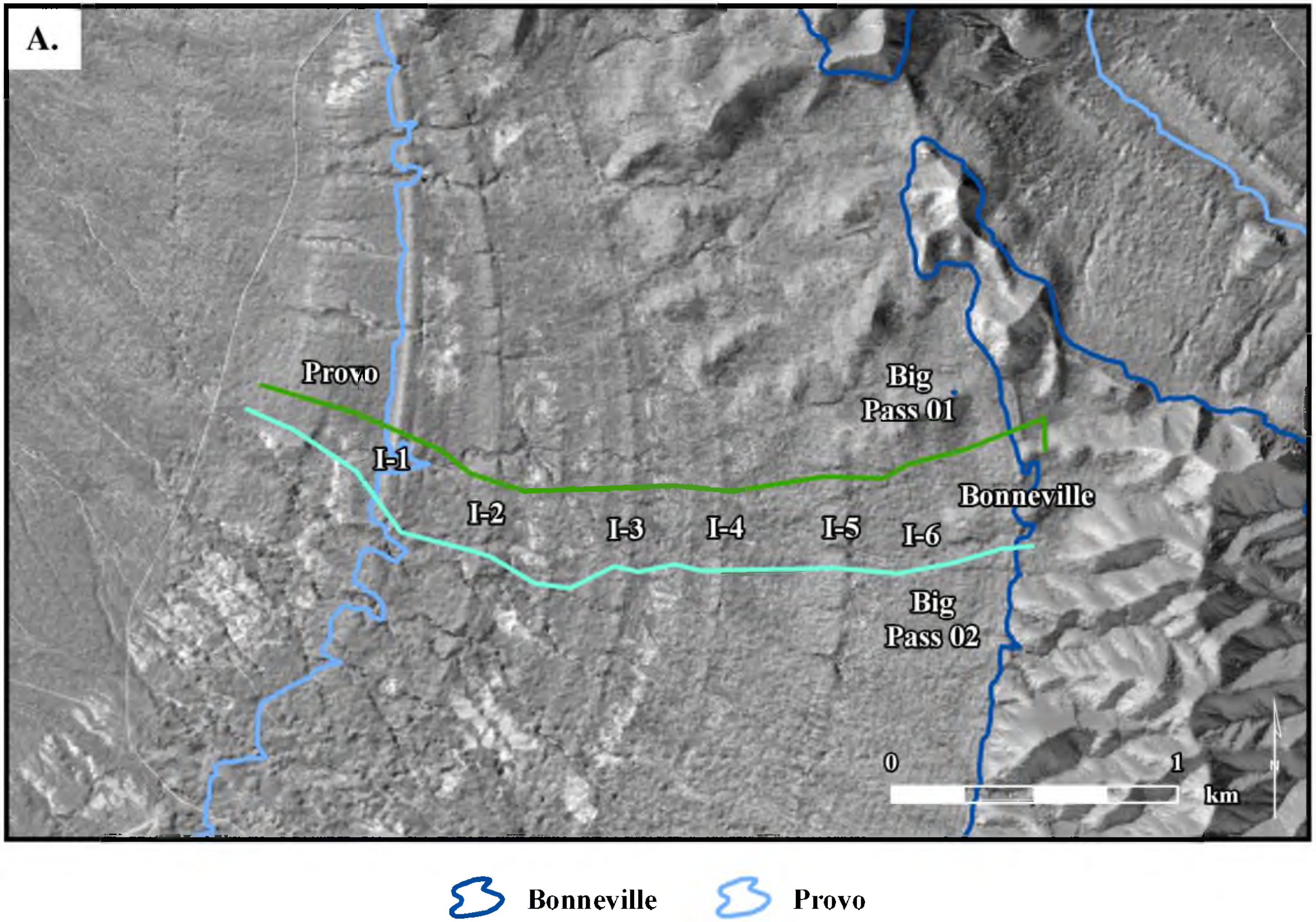


Figure 4.11: Big Pass altitudinal profiles corrected with the ILM. Profiles start slightly above the Bonneville level and end below the Provo level. A) Aerial photography and profile locations, B) Attitudinal profiles.

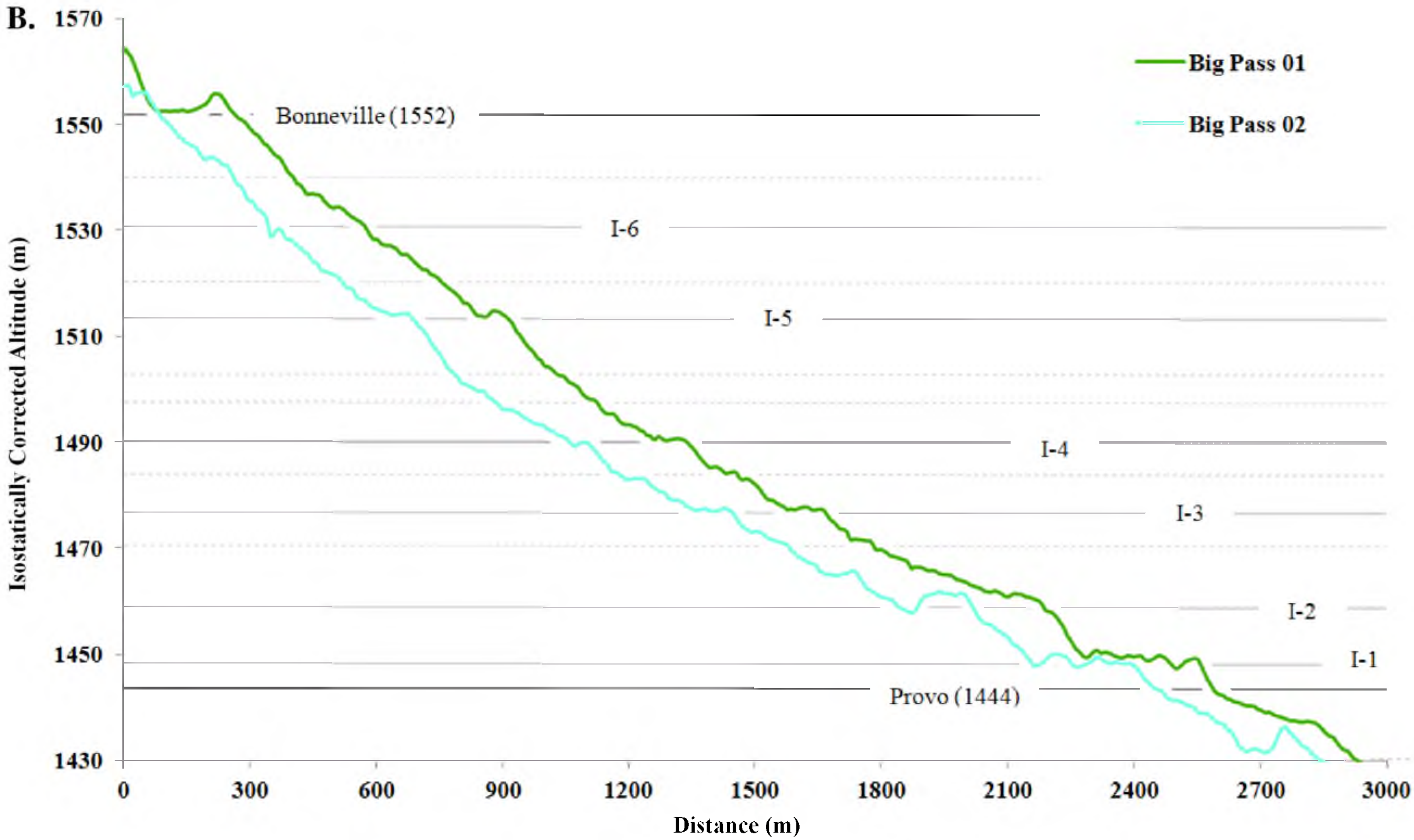
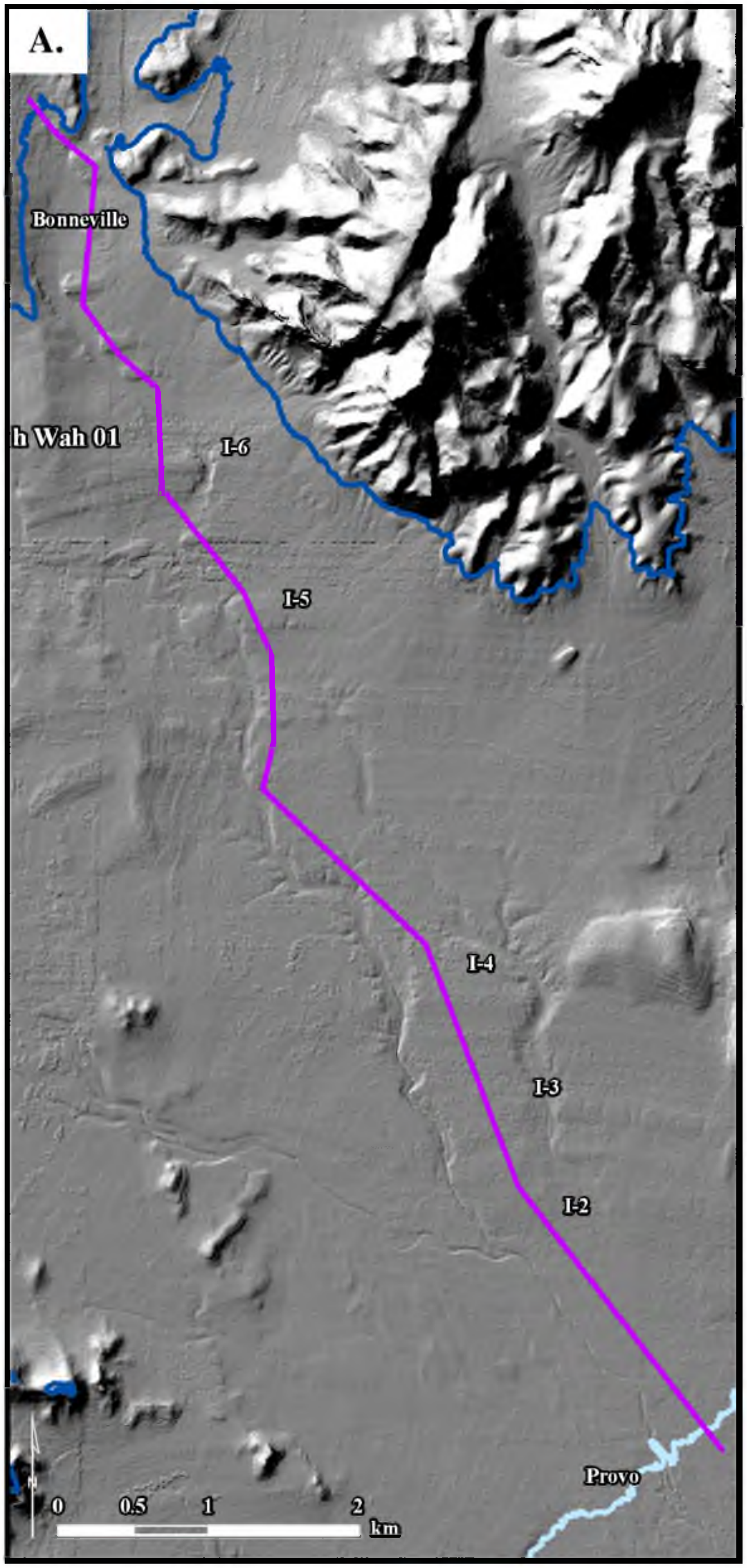


Figure 4.11 (cont.)

Figure 4.12: Wah Wah Valley west group altitudinal profiles corrected with the ILM. Profiles start slightly above the Bonneville level and end below the Provo level. A) Aerial photography and profile locations for Wah Wah 01, B) Aerial photography and profile locations for Wah Wah 02, C) Attitudinal profiles.



Bonneville Provo

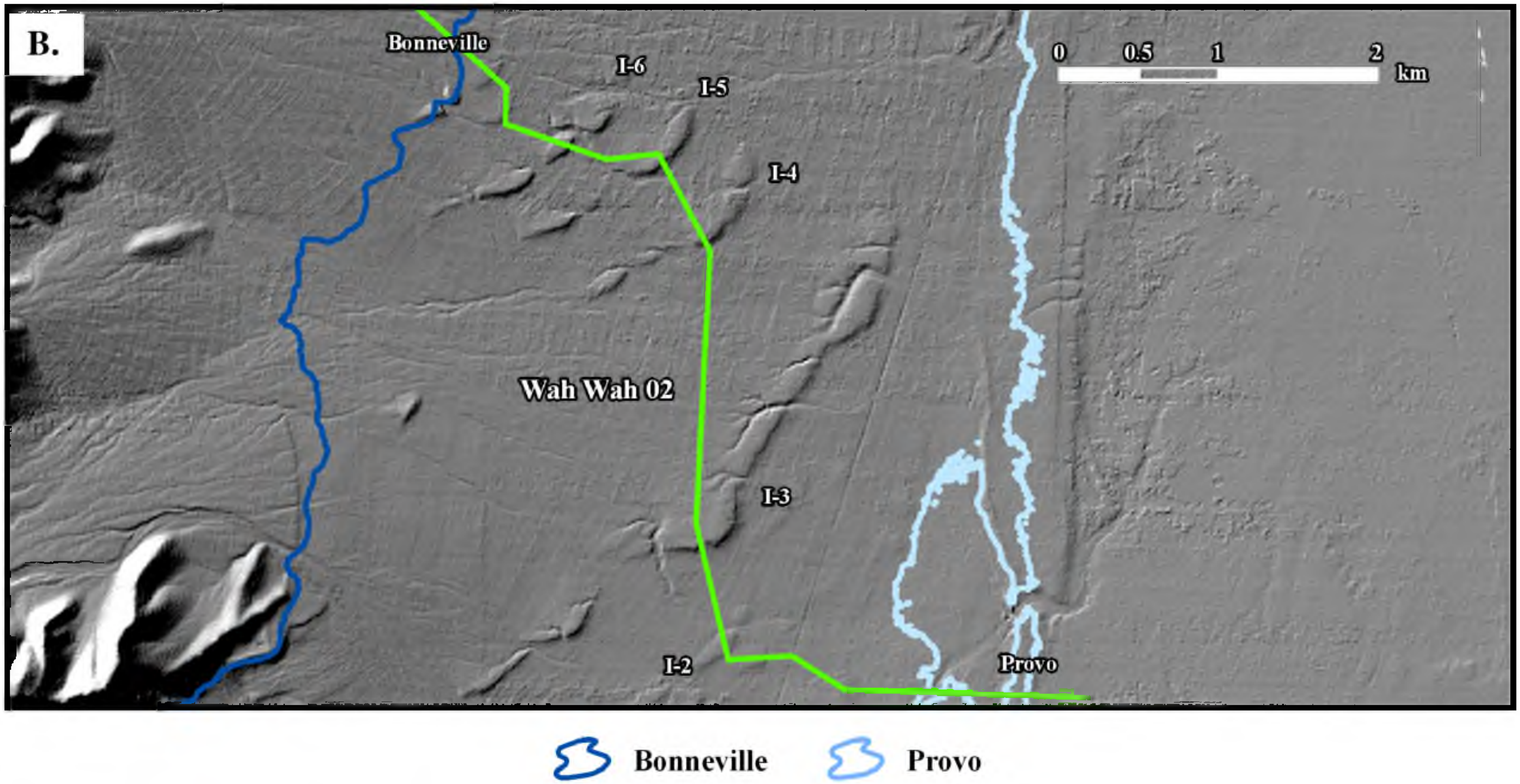


Figure 4.12 (cont.)

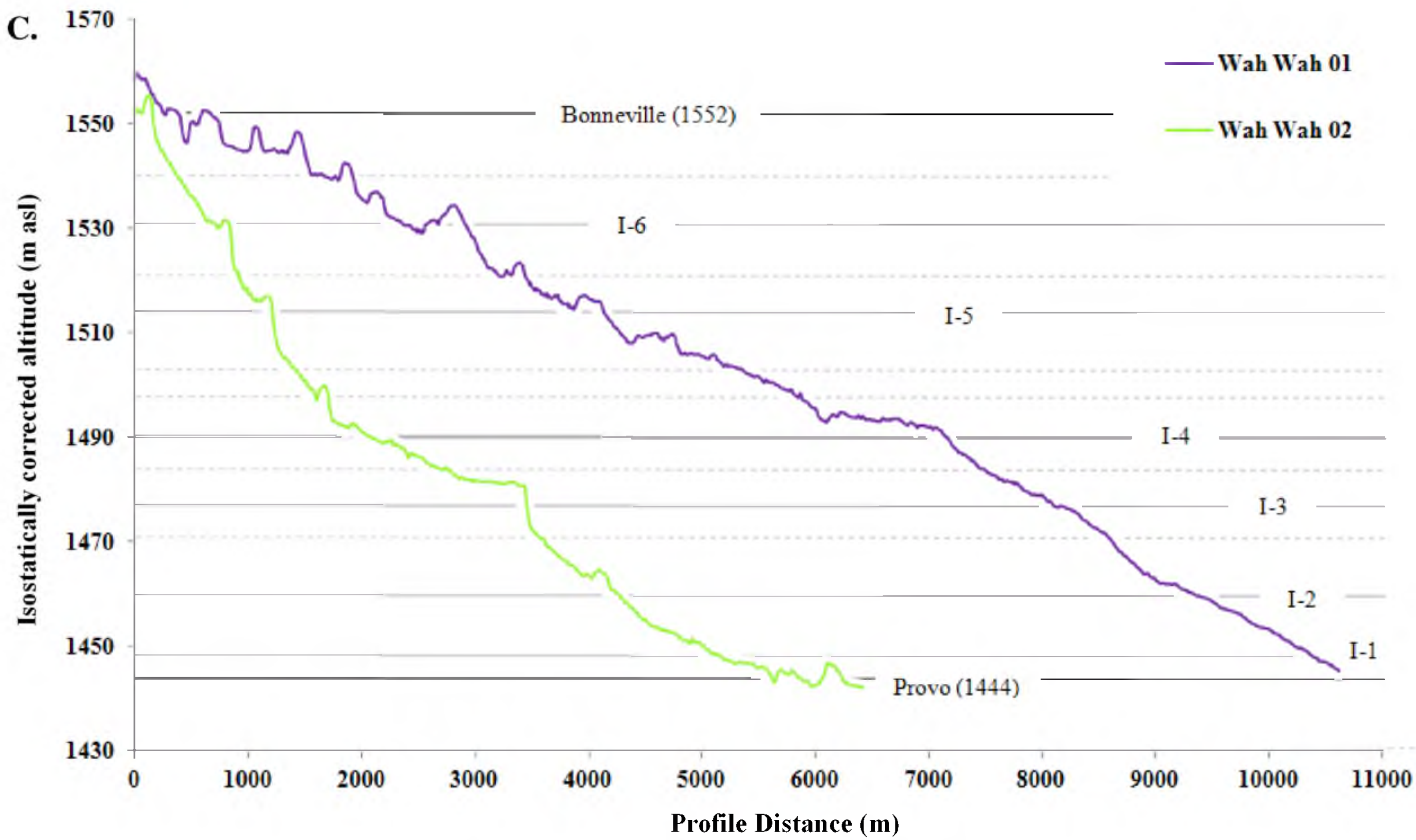


Figure 4.12 (cont.)

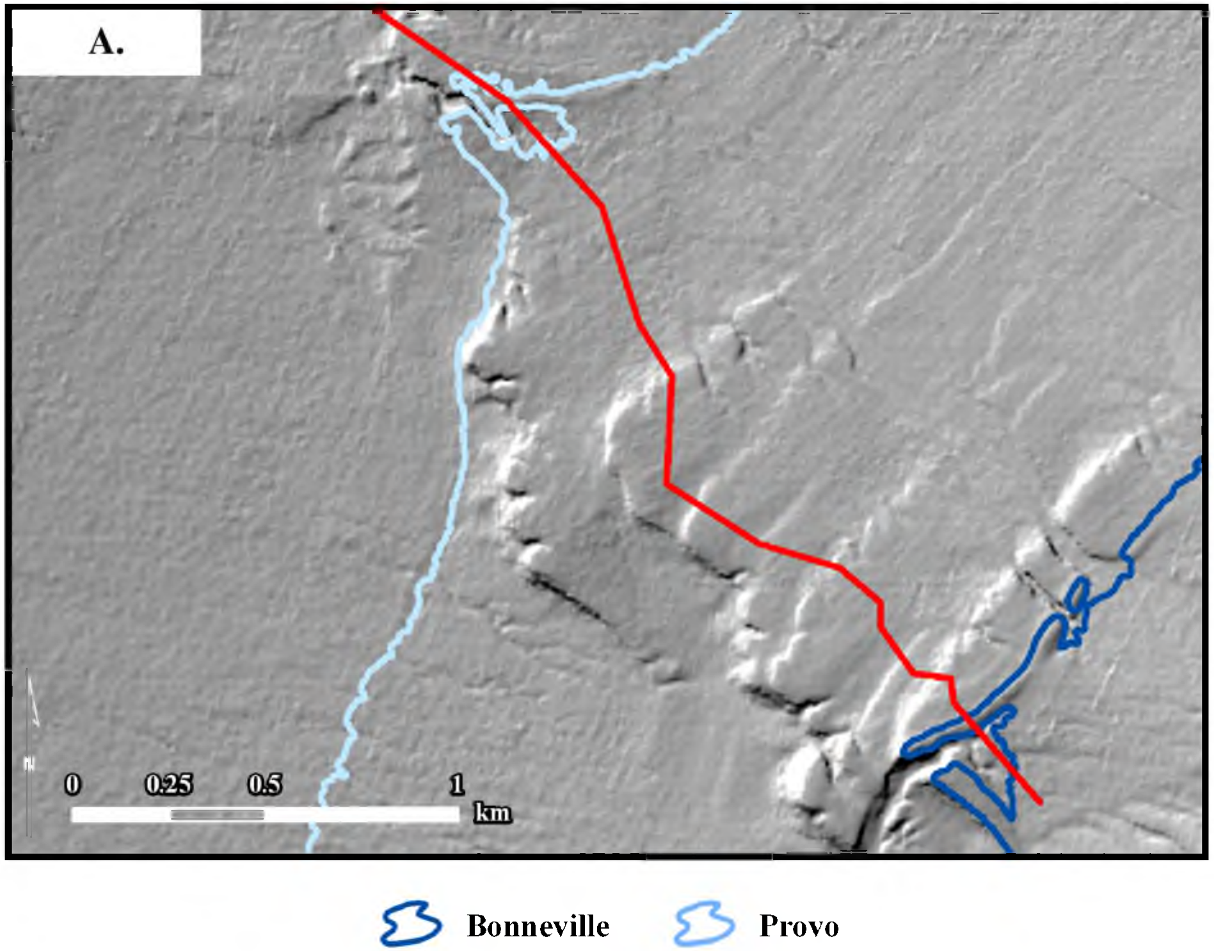


Figure 4.13: Wah Wah Valley east group altitudinal profiles corrected with the ILM. Profiles start slightly above the Bonneville level and end below the Provo level. A) Aerial photography and profile locations for Wah Wah 03, B) Aerial photography and profile locations for Wah Wah 04, C) Attitudinal profiles.

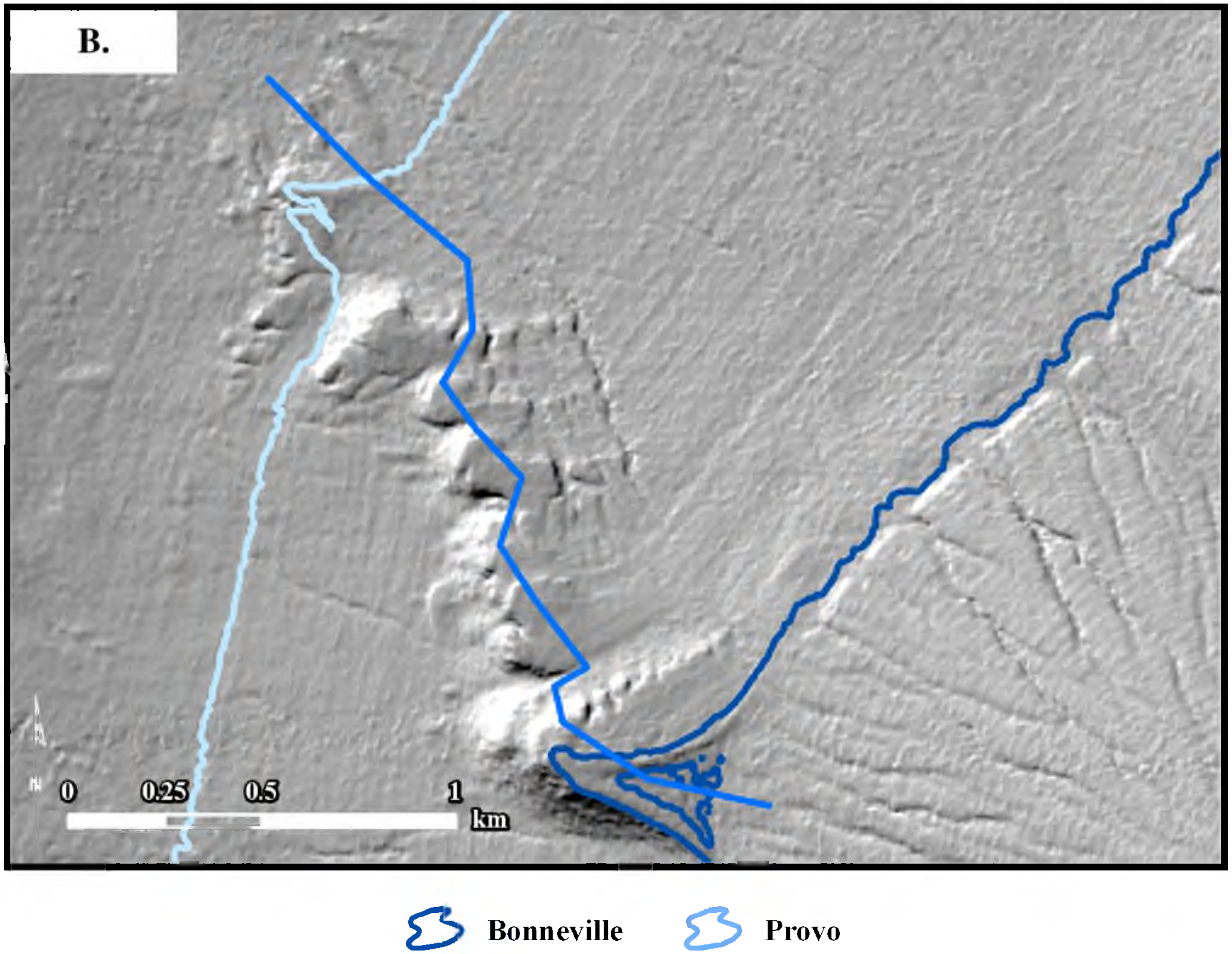


Figure 4.13 (cont.)

C.

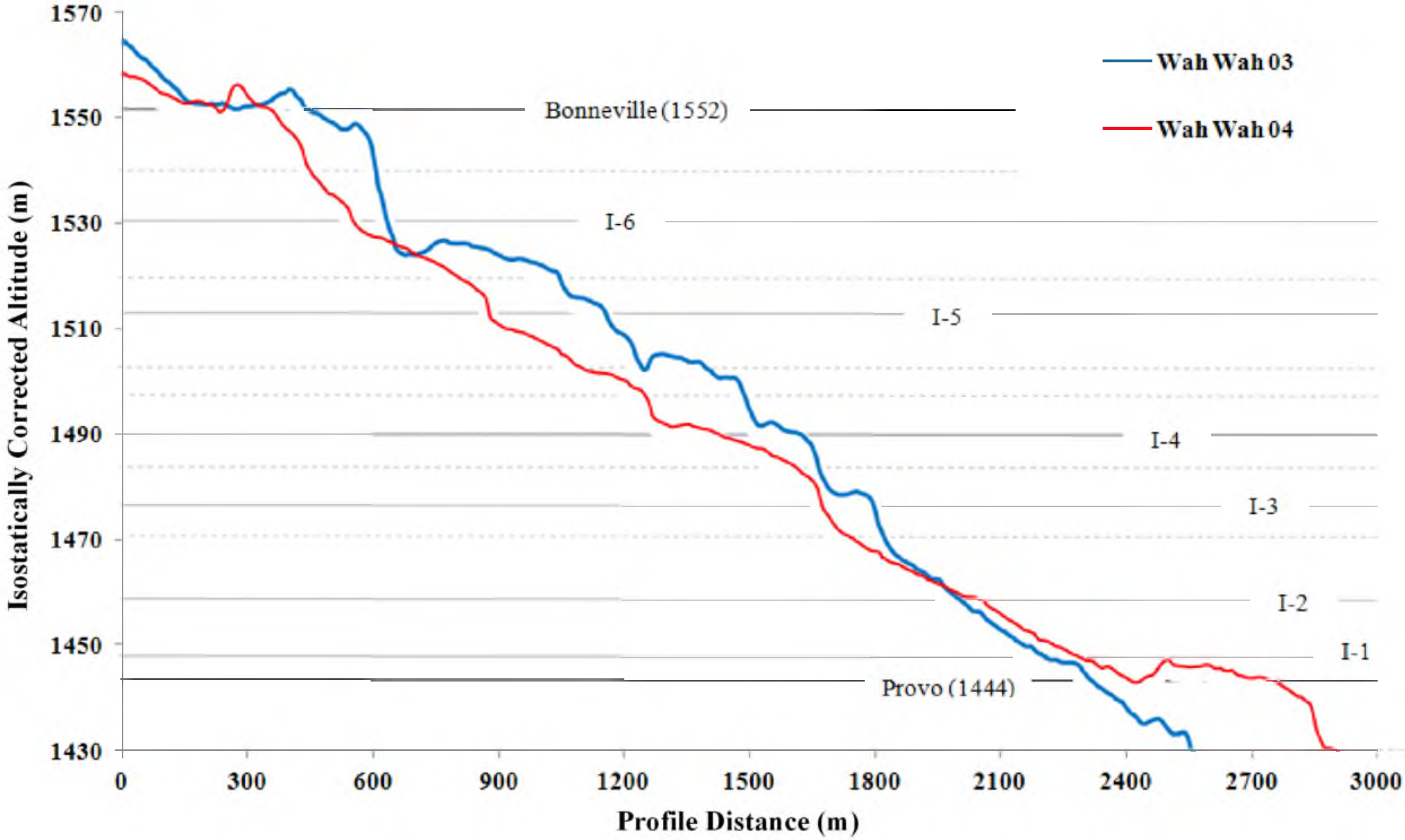
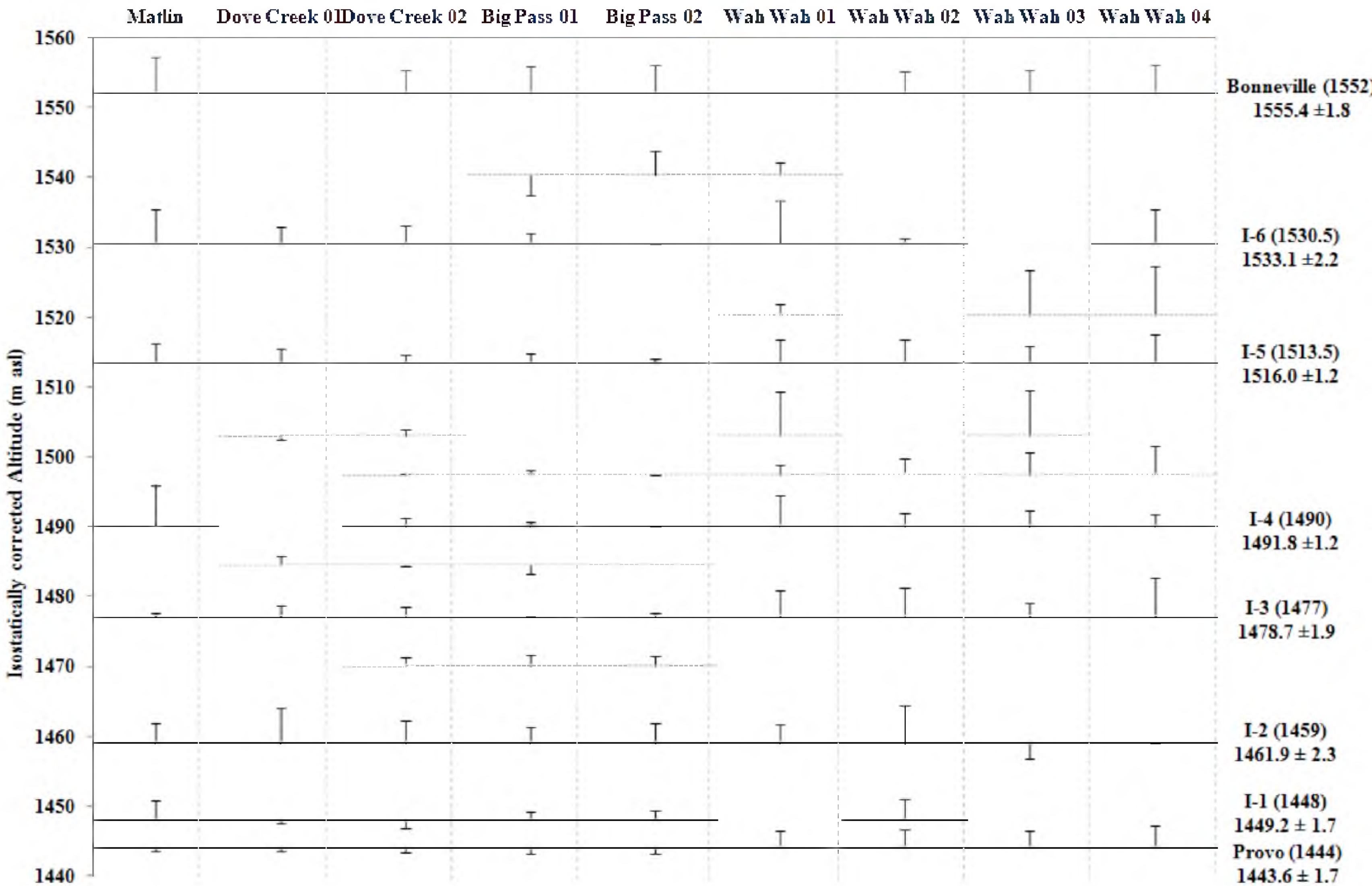


Figure 4.13 (cont.)

Figure 4.14: Variance of the correlated profile altitudes from the interpreted SWL. Solid lines represent SWL of significant paleoshorelines and dashed lines represent other correlated paleoshorelines. The lines are omitted if the visual expression of the paleoshoreline was not visible in the profiles. The bars represent the variance from the SWL to either the crest or erosional notch of the paleoshoreline profile. The median altitude and standard deviation of the variance are recorded for each of the significant paleoshorelines.


Matlin
Dove Creek 01
Dove Creek 02
Big Pass 01
Big Pass 02
Wah Wah 01
Wah Wah 02
Wah Wah 03
Wah Wah 04
Isostatically corrected Altitude (m asl)
1560
1550
1540
1530
1520
1510
1500
1490
1480
1470
1460
1450
1440
Bonneville (1552)
1555.4 ±1.8
I-6 (1530.5)
1533.1 ±2.2
I-5 (1513.5)
1516.0 ±1.2
I-4 (1490)
1491.8 ±1.2
I-3 (1477)
1478.7 ±1.9
I-2 (1459)
1461.9 ± 2.3
I-1 (1448)
1449.2 ± 1.7
Provo (1444)
1443.6 ± 1.7

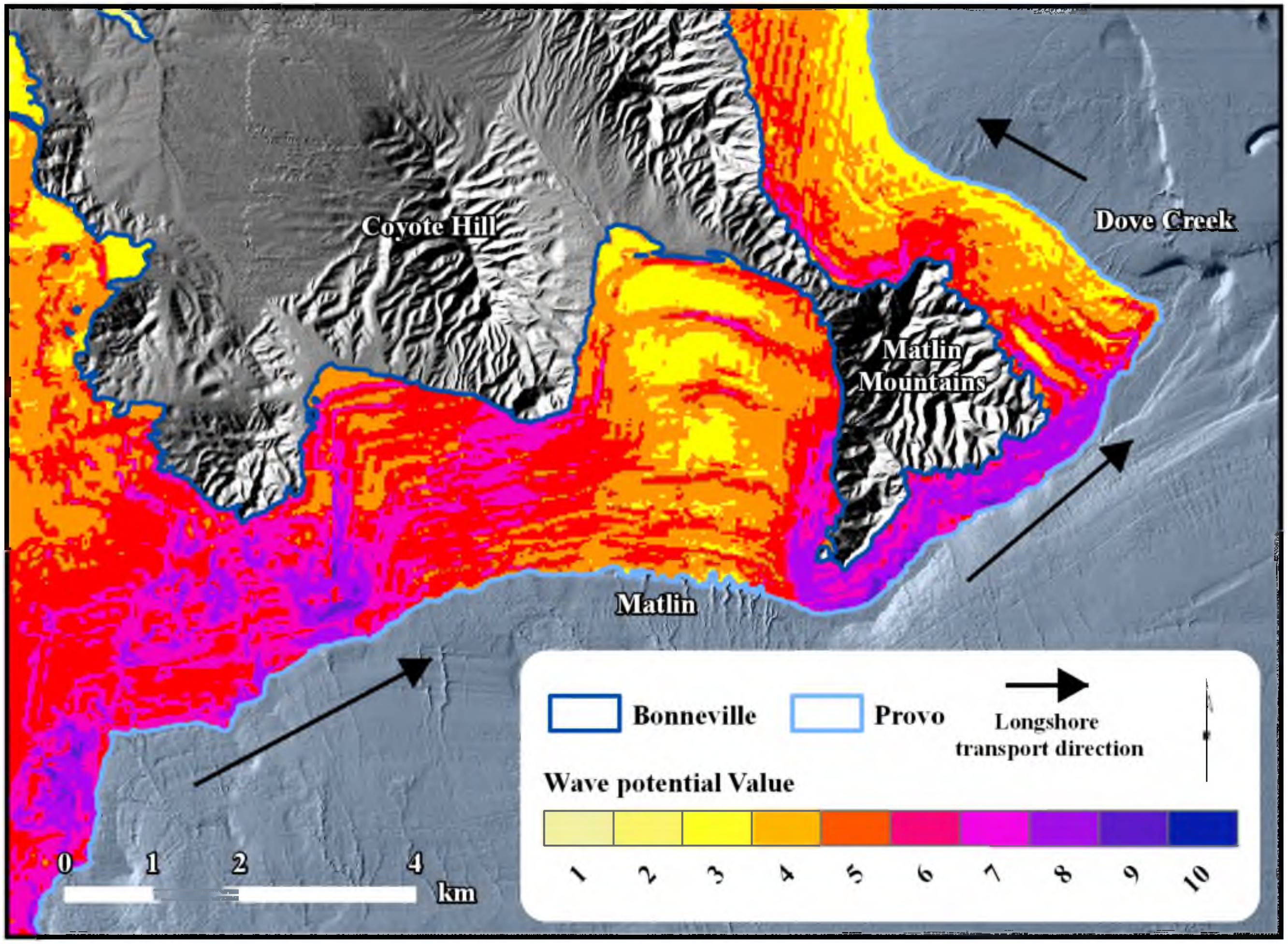
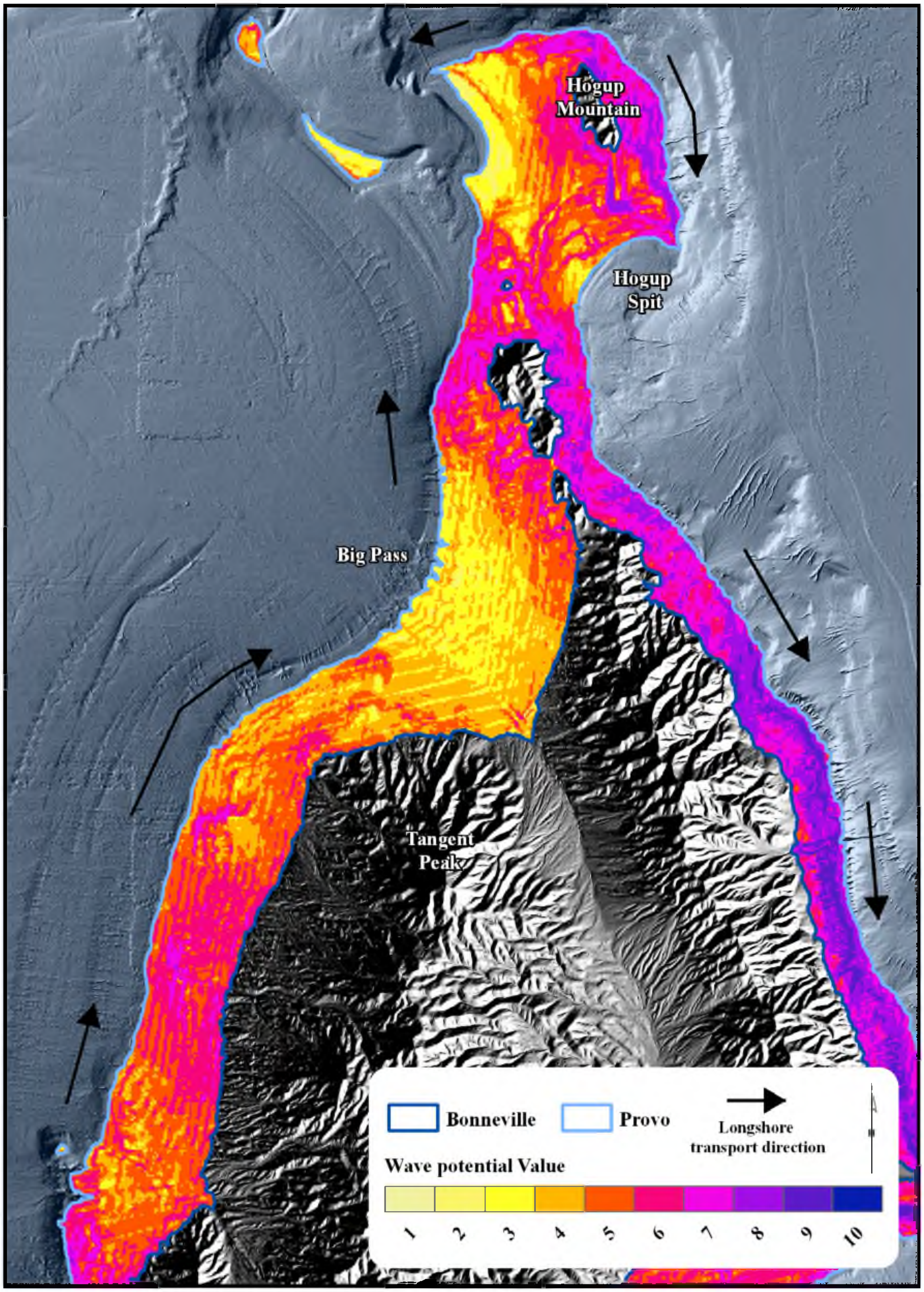


Figure 4.15: Dove Creek and Matlin Mountain Potential Wave Energy Model (PWEM) raster. The arrows indicate the prominent direction of longshore sediment transport delineated from sedimentological and geomorphic evidence.

Figure 4.16: Big Pass Potential Wave Energy Model (PWEM) raster. The arrows indicate the prominent direction of longshore sediment transport delineated from sedimentological and geomorphic evidence.


Hogup Mountain
Hogup Spit
Big Pass
Tangent Peak
Bonneville
Provo
Longshore transport direction
Wave potential Value
1
2
3
4
5
6
7
8
9
10

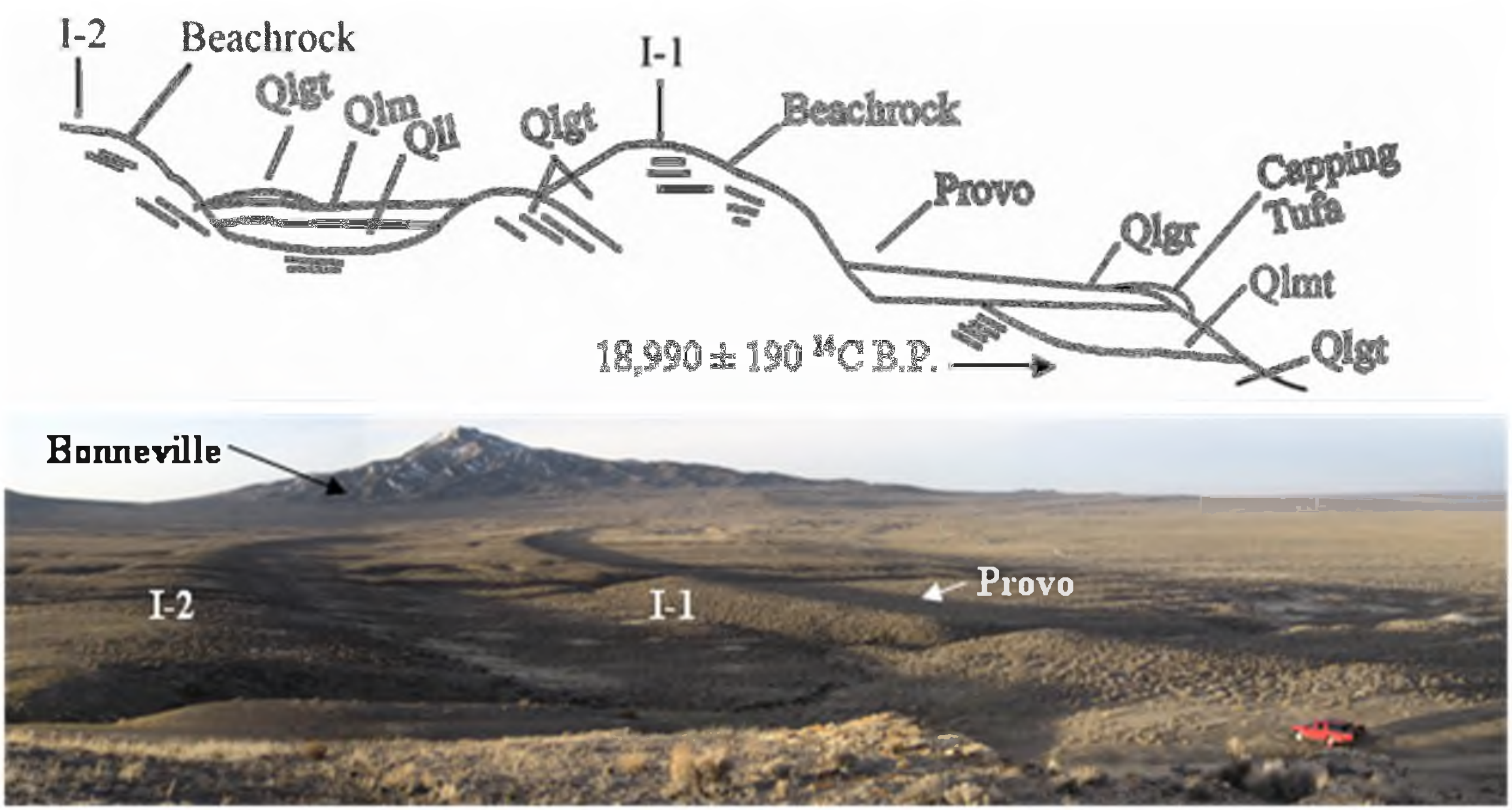


Figure 4.17: Relationship between I-1, I-2, and Provo deposits at Big Pass. Locally the Provo paleoshoreline is an erosional features that eroded into the older I-1 paleoshoreline ridge. Qlgt: transgressive age lacustrine gravels, Qlmt: transgressive age fine grain lacustrine sediments (marls and calcareous rich sand, silt and clay), Qll: lacustrine lagoonal muds and sands, Qlgr: regressive lacustrine gravels (Provo age), Beachrock: carbonate cemented gravels, and capping tufa deposits. Radiocarbon sample was measured from gastropod sample collected from transgressive offshore lacustrine sands (Sack, 1999). Picture taken from north side of the embayment at Big Pass with Tangent Peak in the background. Red truck for scale.

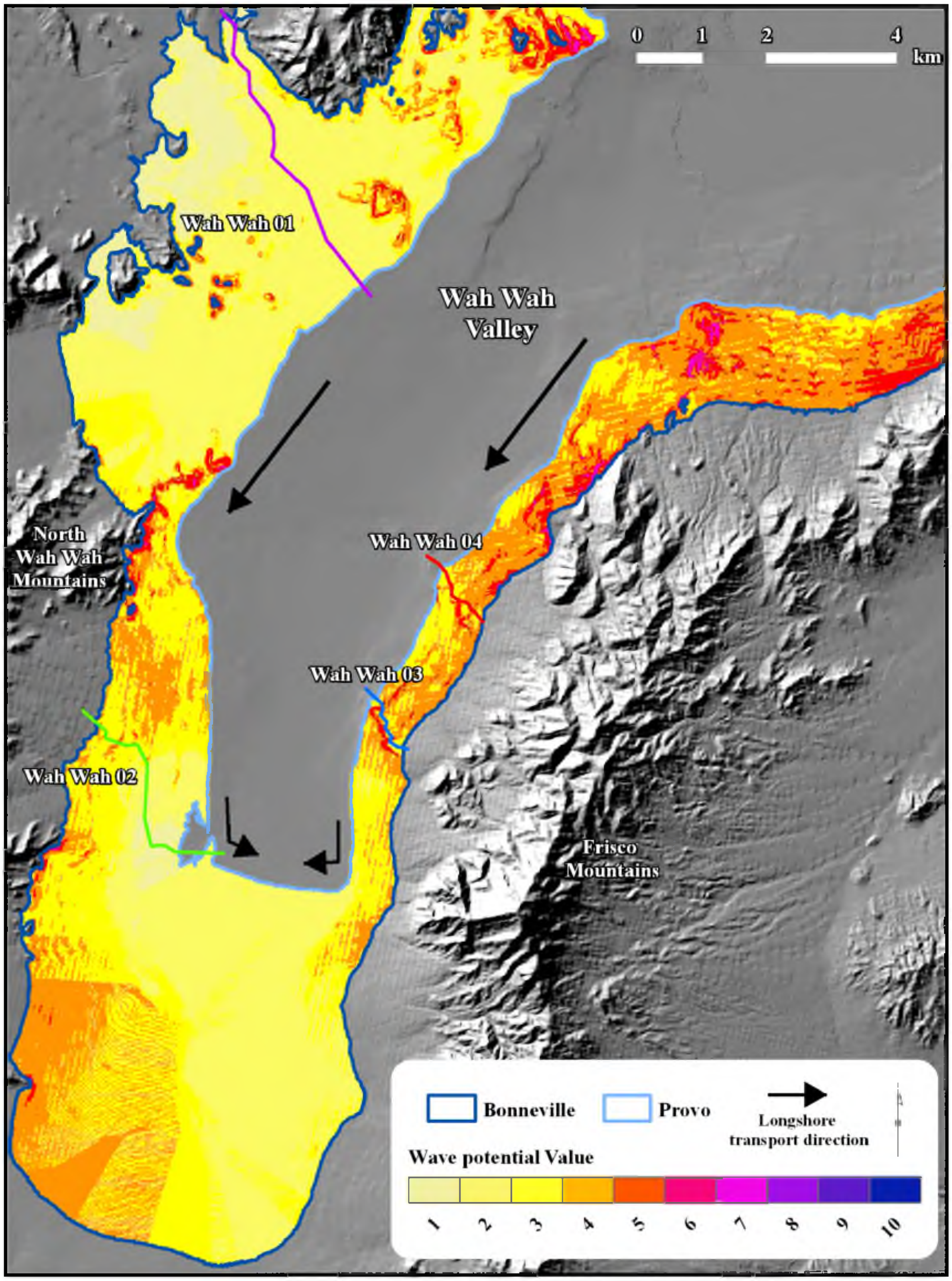


Figure 4.18: The altitude profile locations for the Wah Wah Valley and the Wah Wah Valley Potential Wave Energy Model (PWEM) raster. The arrows indicate the prominent direction of longshore sediment transport delineated from sedimentological and geomorphic evidence.

Figure 4.19: A. Generalized hydrograph of Lake Bonneville modified from Oviatt (1997) and Godsey et al. (2011). B. Plot of age and isostatically corrected altitudes of samples in Table 5.1 (Nelson and Jewell, in review) in relation to the Lake Bonneville hydrograph. Amplitude limits of water level fluctuations associated with the U1, U2, and U3 oscillations are approximate and shown here schematically. Open circles indicate previously published data, whereas closed circles represent data from Nelson and Jewell (in review). The horizontal lines emanating from the sample points represent the error bars related to the age of the sample.

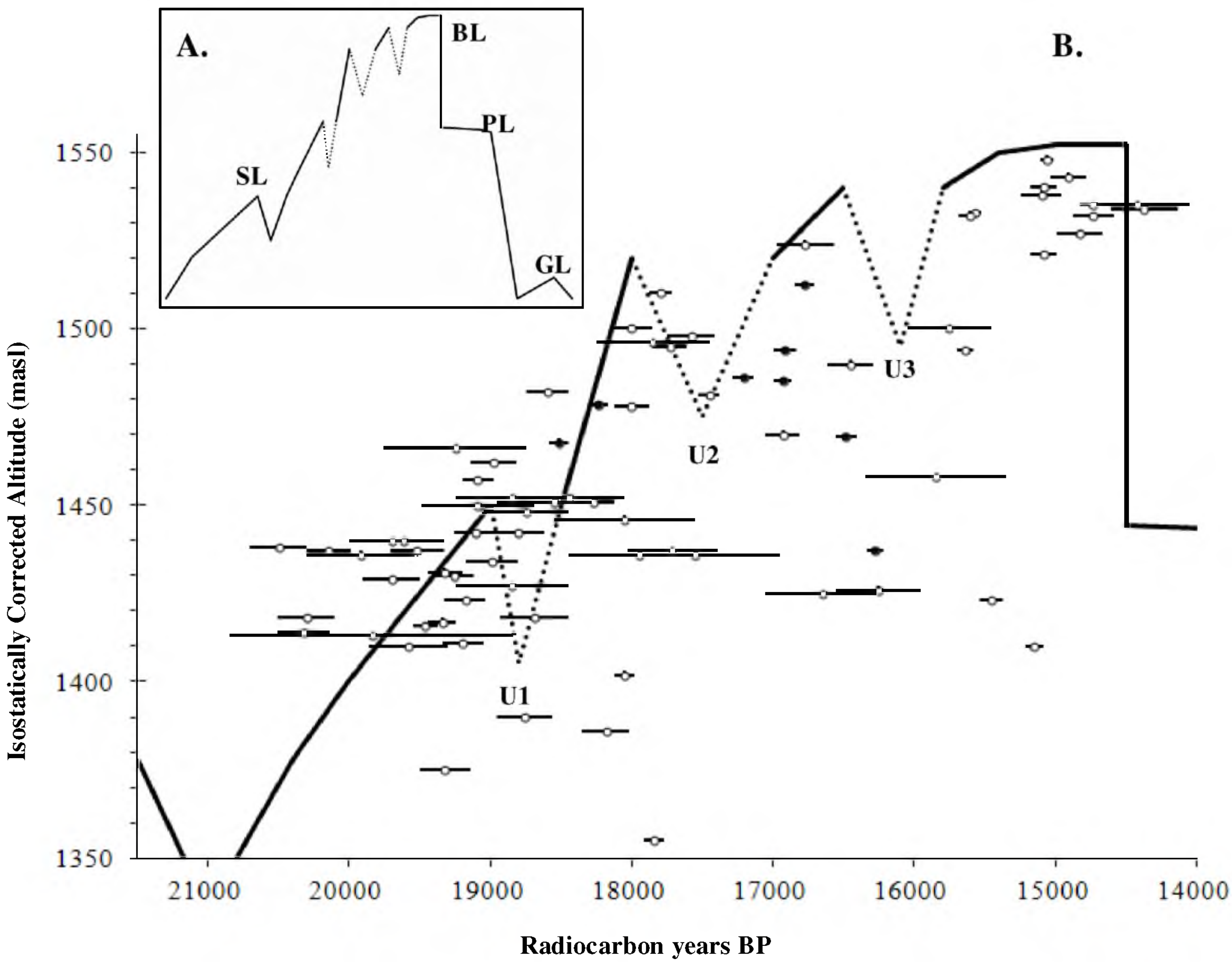

A.
SL
BL
PL
GL
B.
U1
U2
U3
Isostatically Corrected Altitude (masl)
1350
1400
1450
1500
1550
Radiocarbon years BP
21000
20000
19000
18000
17000
16000
15000
14000

Table 4.1: Measured altitudes of local Bonneville and Provo still water levels (SWL).

Isostatic correction at each of paleoshorelines, and the differences of the values. Altitudes are measured in meters above sea level and isostatic correction is measured in meters.

	Hogup Spit	Big Pass	Matlin	Dove Creek	S WahWah	N WahWah	Snowplow	Grantsville
Bm	1593	1602	1592	1592	1557	1558	1589	1600
Bc	41	50	40	40	5	6	37	48
Pm	1476	1481	1476	1476	1451	1454	1472	1484
Pc	32	37	32	32	7	10	28	40
Bm - Pm	117	121	116	116	106	104	117	116
Bc - Pc	9	13	8	8	-2	-4	9	8

Bm	Locally measured altitude of Bonneville paleoshoreline
Bc	Isostatic correction for Bonneville: (Bm – 1,552)
Pm	Locally measured altitude of Provo paleoshoreline
Pc	Isostatic correction for Provo: (Pm – 1,444)

Table 4.2: Divergence of the differing hydro-isostatic rebound methods.

Divergence of the differing hydro-isostatic rebound methods from the ILM at the Bonneville and Provo still water levels from selected altitude profiles.

	Wah Wah S		Wah Wah Mid		Hogup Spit 01		Hogup Spit 02	
	Provo	Bonneville	Provo	Bonneville	Provo	Bonneville	Provo	Bonneville
BLM	5.62	-0.01	5.62	-0.01	2.40	0.00	2.40	0.00
ISM Bonneville	3.94	0.53	2.81	0.55	-12.43	3.03	-12.52	0.06
ISM Provo	2.16	-2.21	1.04	-3.01	-3.36	9.39	-1.80	9.24

Table 4.3: Reclassification scheme of fetch values.

Reclassification scheme and weighting of maximum slope, maximum fetch, and median fetch. The reclassification scheme of the fetch values were based on a Jenks natural breaks algorithm.

Reclassification Value	Maximum Slope (Degrees)	Maximum Fetch (km)	Median Fetch (km)	
1	0-1	0-26	0-4	
2	1-2	26-47	4-8	
3	2-3	47-69	8-11	
4	3-4	69-92	11-16	
5	4-6	92-113	16-21	
6	6-8	113-135	21-28	
7	8-10	135-159	28-36	
8	10-15	159-186	36-45	
9	15-25	186-218	45-58	
10	25-90	218-340	58-97	
Weight	40%	30%	30%	100%

CHAPTER 5

THE OSCILLATORY RECORD OF LATE PLEISTOCENE TRANSGRESSIVE PALEOSHORELINES OF LAKE BONNEVILLE, U.S.A.

Abstract

New stratigraphic and chronological data within the lacustrine sediments of the Hogup Mountains of northwestern Utah provide evidence for two previously unpublished oscillations and further evidence for another previously proposed oscillation in the transgressive record of late Pleistocene Lake Bonneville. Since the lake's level is closely tied to climatic responses, resolving stratigraphic and chronologic history has the potential to improve the resolution of regional and global paleoclimate models. The sedimentological deposits at this site are related to the transgressive paleoshorelines between the altitudinal limits of the Bonneville and Provo levels (Intermediate paleoshorelines). Including the two newly proposed oscillatory events from this paper, there are seven proposed oscillatory events during the transgressive phase (25 – 14.5 ^{14}C kyr B.P.) of Lake Bonneville: two Stansbury oscillations (~21.5 and 20.5 ^{14}C kyr B.P.), the U1 oscillation (~18.6 ^{14}C kyr B.P.), the newly proposed Lower Hogup Oscillation (~18.2 ^{14}C kyr B.P.), the U2 oscillation (~17.5 ^{14}C kyr B.P.), the newly proposed Upper Hogup Oscillation (17.2 ^{14}C kyr B.P.), and the U3 oscillation (~16.4 ^{14}C kyr B.P.). It is

proposed that these Intermediate oscillations record submillennial oscillations in the lake level that likely correlate to short term climate shifts during the Last Glacial Maximum.

Introduction

During the late Pleistocene at the height of the Last Glacial Maximum, Lake Bonneville was the largest pluvial lake of the western United States (Fig. 5.1). Widespread sedimentological and geomorphic paleoshoreline features formed in the basin as the lake's hydrologic budget, resultant water level, and hydrodynamic processes continually changed with the dynamic climate of the late Pleistocene. The general hydrologic chronology of the lake's four major paleoshorelines (i.e., Stansbury, Bonneville, Provo, and Gilbert lake levels—Fig. 5.2) has been well established and directly related to the region's broad climatic history (Oviatt, 1997; Kaufman, 2003; Balch, 2005; Godsey et al., 2005; 2011; Oviatt et al., 2005). However, there is still little understanding of past submillennial climatic trends in the basin.

In addition to the four major paleoshorelines many less prominent paleoshoreline features are found within the basin. In his 1890 Monograph, G.K. Gilbert discussed a subset of these less prominent paleoshorelines that he termed the "Intermediate shore lines" which in this paper will be referred to as "Intermediate paleoshorelines" and the time in which they formed will be called "the Intermediate time" or "Intermediate period." Gilbert identified the Intermediate paleoshorelines as lacustrine landforms formed between the altitudinal limits of the Bonneville and Provo paleoshorelines. Gilbert identified multiple localities in the basin where Intermediate paleoshorelines were expressed (e.g., Dove Creek, Wah Wah Valley, the Old River Bed, Stockton Bar, Grantsville spits [South Willow]—Fig. 5.1) and discussed possible explanations for the variation seen in the altitude of the features. Even though other researchers mention the existence of these less prominent features (Scott et al., 1983; Scott, 1988; Burr and Currey, 1992; Oviatt et al., 1994; Sack, 1999) and periodically included them in regional geologic maps (Miller & Oviatt, 1994; Miller & McCarthy, 2002), the literature does not describe their relevance in much detail.

We hypothesize that stratigraphic evidence related to these Intermediate paleoshorelines record a sub millennial oscillating lake record superimposed on the broader climatic trends inferred from the more general paleoshoreline record. The Intermediate paleoshorelines are important because they were deposited during the height of the Last Glacial Maximum (LGM) as the region's climate was responding to global climatic shifts (Fig. 5.2). Oviatt (1997) has proposed that the transgressive lake recorded multiple oscillations (U1 – U3) in lake level directly related to global climatic events such as termination of Heinrich events (Oviat, 1997; Benson et al., 1997; 2003; 2011), Dansgaard-Oeschger events (Benson et al., 1997; 2003; 2011), and other global manifestations related to fluctuations of the Laurentide Ice Sheet.

Deposits related to the lake level oscillations during formation of the Intermediate paleoshorelines, have only been stratigraphically described with radiometric context in the near shore deposits of the Sevier subbasin (Oviatt, 1997). The purpose of this study is to (1) provide additional sedimentological evidence and age control for previously and newly proposed oscillatory events during the Intermediate period of Lake Bonneville's transgressive phase; and (2) discuss how these proposed oscillatory events may relate to climatic changes during and immediately following the Last Glacial Maximum.

Previous Studies

Gilbert (1890) observed that individual Intermediate paleoshorelines have significant altitudinal differences in their geomorphic expressions (e.g., barrier ridges, spits, wave platforms). In addition, he observed that the same number and/or size of Intermediate paleoshorelines are not preserved from location to location. Gilbert suggested that hydro isostatic adjustment of the Bonneville basin, differential distribution of wave energy, and the autogenic (internally driven) sedimentological effects of an oscillating lake level could have been some of the factors responsible for the variations seen in the Intermediate paleoshorelines. However, Gilbert concluded that the autogenic sedimentological effects of an oscillating lake was a dominant control on the altitude and preservation variations observed in the Intermediate paleoshorelines.

During the transgressive phase of the Lake Bonneville cycle, the basin was hydrologically closed. The lake's water budget and resultant water level were adjusting to both seasonal variations in weather patterns and long term climatic shifts. Currey and Oviatt (1985) initially suggested multiple oscillations that they termed "stillstands" that were times when the lake surface stabilized due to the flow of the lake into internal sub basins during the transgressive phase of the lake. However, the evidence for most of these oscillations and stillstands were not well documented. Oviatt (1997) further suggested that there were up to five large oscillations of lake level during the 27,000 to 15,000 ^{14}C yr B.P. transgressive rise of the lake. The Stansbury Oscillation(s) (20,000–22,000 ^{14}C yr B.P.) is the most prominent and well studied of these transgressive oscillations and may have consisted of two distinct oscillations (Oviatt, 1987; Oviatt et al., 1990; Patrickson et al., 2010; Benson et al., 2011). Oviatt (1997) also suggested up to three (U1 – U3) more large oscillations (amplitudes of ~45 m) during the Intermediate history of the lake from 20,000 to 15,000 ^{14}C yr B.P (Fig. 5.2). These Intermediate lake level oscillations (U1-U3) are inferred from sedimentation patterns of gravel wedges found in outcrops within the lake's Sevier subbasin and from $\delta^{18}O$ patterns in a few basin sediment cores (Oviatt 1997). Well-dated stratigraphic contexts of the oscillations (U1-U3) have not been documented at other localities around the paleolake's relict shore face deposits.

The Hogup Mountain Locality

This paper describes the well exposed stratigraphy of the lake's Intermediate landforms within multiple ephemeral gullies on the eastern flanks of Hogup Mountain (Fig. 5.3) in the northwestern portion of the Bonneville Basin. Geomorphic and sedimentological inferences (i.e., direction of spits, sediment size patterns and sources) for longshore transport during the transgressive period, suggest that the sediment sources for many of these transgressive landforms are from the reworking of pre-Bonneville alluvial fans or excavated from the fractured bedded limestone and quartzite deposits of the Permian and early Pennsylvanian Oquirrh Group (Stifel, 1964; Doelling, 1980).

Methods

The northern Hogup Mountains have a well-preserved record of the four major paleoshorelines, Intermediate paleoshorelines, and many other minor paleoshorelines deposited in both the regressive and transgressive phases of the lake. The paleoshorelines for the area were mapped based on 1:10,000 scale aerial photography. This paper describes the stratigraphic and geomorphic relationships of the deposits within the ephemeral gullies and barrier ridges of the Hogup Mountains. Stratigraphic descriptions concentrate on sedimentary trends of multiple properties (i.e., grain size, composition/provenance, and degree of rounding), transitions between beds (unconformities or continued deposition), sedimentary structures, and the abundance and condition of fossil material (i.e., gastropods and ostracodes). The geomorphic analysis of the landforms in the area was accomplished by on site field investigations, aerial photo and topographic map interpretation, analysis of digital elevation models, and other Geographic Information System (GIS) techniques. Unless specifically stated, all altitudes reported in this paper have been corrected for isostatic rebound based on methodology used by Oviatt et al. (1992) and adapted by Nelson (2012). The methodology simplifies the complexities of isostatic adjustment by projecting these corrections as a linear approximation of the altitude changes due to isostatic adjustment between the Provo and Bonneville lake stages.

Freshwater gastropod shells (*Stagnicola Bonnevillensis*) were collected and analyzed for radiocarbon ages to provide a chronological constraint on the ages of the deposits and proposed oscillatory events. The sediment in which the shells were collected was thoroughly examined for indications of reworked shells (i.e., broken ostracod and gastropod shells within the sediments), and only intact whole gastropod shells were submitted for analysis. The shells were analyzed by accelerator mass spectrometry (AMS) by Beta Analytic Inc. in Miami, Florida. The samples were washed with a mild acid prior to analysis. In order to be consistent with other Lake Bonneville researchers, ages are reported in radiocarbon yrs B.P. The reported ages of the Hogup Mountain samples are compared to other published ages for the occupation of the Intermediate paleoshorelines (Fig. 5.3) and reported in Table 5.1. Consistent with the methodology of

Godsey et al. (2005), 350 yrs were added to ages measured prior to 1977 in order for the ages to be consistent with modern reporting standards (Stuvier and Polach, 1977). The samples were not corrected for hard water effects or post depositional contamination by younger carbon because the adjustments for these effects are assumed to be within analytical errors (Benson and Thompson, 1978; Oviatt et al., 1992; Godsey et al., 2005). In order to be consistent with the reporting of samples ages, ages that have previously been adjusted for hard water effects (i.e., + 500 yrs, Broecker and Kaufman, 1965) have been adjusted back to their original reported age (Scott et al., 1983; Oviatt et al., 1992; Godsey et al., 2005).

Results

Multiple ephemeral gullies have been produced by surface runoff in the study area since the late Pleistocene. This runoff has exposed the internal stratigraphy of lacustrine sediments draped over the fractured bedrock of the Oquirrh Group. The composite profile of these sediments (Fig. 5.4) are compiled from multiple measured sections of exposed lacustrine sand, gravel, and marl units at sample locations 19, 27, 28, 43, and 46 (Fig. 5.3). Multiple wave cut bedrock platforms on the northern flanks of the Hogup Mountain were the sediment source area for the depositional features on the western and southern flanks of the mountain. The lacustrine sediments in the region are relatively thin, but the deposits thicken and the sediments get finer and more mature in a southerly direction. This lateral sedimentation pattern suggests a southerly direction of longshore transport. These clastic deposits have the geomorphic expression of barrier ridges buried by offshore deposits as the lake transgressed to the Bonneville level.

Three marl units are present within the composite profile (Fig. 5.4): the lower marl (LM), the middle marl (MM), and the upper marl (UM). Each marl units is positioned between lacustrine sand and gravel units: the basal gravel (BG), lower gravel (LG), middle gravel (MG), and upper gravel (UG). All three marl deposits are interstratified with the gravel and sand units (Fig. 5.4).

The basal gravel (BG) drapes over the original bedrock surface of the slope. The gravel is a well sorted, well rounded, medium gravel with no visible sedimentary structures. At the top of the gravels, the deposit conformably grades into a very thin (~6 cm) grey sand with multiple ostracodes and gastropods (*Stagnicola Bonnevillensis*). A gastropod sample from this horizon (sample 46, Table 5.1) has an age of 18,510 ^{14}C yr B.P. Even though some fragmented ostracodes can be found, most are preserved as a half shell or as a whole (paired) sample; therefore, the preservation and condition of these shells suggest the material has not been significantly reworked.

Conformably overlying the fine sand lens of the BG is the lower marl (LM). The LM can be separated into two sections. The lower section, composed of a sandy marl with multiple fine sand laminations grades upward into an upper layer of clay rich marl. The lower sandy marls dip slightly (~8–10°) shoreward indicating the BG may have been a beach ridge with lower sandy marl sediments draped behind the landform in the backshore; however, the crest of the proposed beach ridge has been eroded away. Sand laminations are ~1–3 cm thick at the base of the unit, but are thin and become less abundant up section. The sandy marls at the base of the unit dip shallowly (~2–6°) shoreward near the fine sand to gravel transition but up section the dip of the sediments gradually shifts basinward.

The lower sand and gravel unit (LG) unconformably overlie the LM deposit. The top of the marl layer is undulated and appears to be truncated or scoured and is followed by the deposition of a fine medium sand or fine medium grained gravel. The fine medium grained, moderately well-sorted, subrounded grey sand is more abundant to the south, and the fine medium gravel unit is more abundant to the north, indicating an erosional surface of higher wave energy followed by the deposition of the landform (i.e., beach ridge) with a southerly longshore transport direction. Both the sand and gravel deposits on lap the marl unit and dip at an angle of 10–15° basinward. There is an abundance of ostracodes (fragmented and half shell) and juvenile gastropods at the base of the sand deposits. A gastropod sample from this horizon (sample 21, Table 5.1) has an age of 16,480 ^{14}C yr B.P. The gravels coarsen upward within the unit, and occasional medium coarse sand lenses are exposed. Near the top of the unit, the gravels dip in a basinward direction and then grade to fine medium sand with ripple laminations (Fig. 5.5).

The middle marl (MM) conformably lies on top of the LG unit as it transitions from ripple laminated sands into a planar laminated sandy marl, followed by a carbonate rich silty marl. At the base of the MM unit, multiple gastropods collected from the fine grained sand yielded (sample 43, Table 1) an age of 18,240 ^{14}C yr B.P. The excellent preservation and condition of ostracodes and gastropods at its base suggest that the sediment from which the sample was collected is not reworked. The middle marl has many pebbles encased within the unit that are interpreted as dropstones.

The silty marl above the sandy marl of MM is truncated and unconformably overlain by down lapping medium grained gravels of the middle gravel (MG) unit. Imbrications in planar laminations of gravels at the base of the unit indicate a north to south transport direction dip basinward (3–10°). Laminated fine medium grained gravels up section within the unit exhibit a steep basinward dip (26–30°). A thin fine grained sand lens (4-6 cm), similar in composition and fossil content to the sediment of sample 43, conformably overlies the gravel unit. This transition between the MG unit and the fine sands of the silty sandy marl is abrupt, and the fine grained sand quickly grades into a thin silty sandy marl (UM) deposit. Gastropods from the fine grained sand deposit (sample 27) yielded an age of 16,930 ^{14}C yr B.P. The upper gravel (UG), a fine medium grained gravel, unconformably on laps both the UM and MG suggesting the marl and gravel were truncated by an erosional event. Another very thin, fine sand with a high abundance of ostracodes and broken gastropods separates the UG from the UM. A gastropod sample (sample 28) from the sand produced an age of 17,210 ^{14}C yr B.P.

Interpretations of the Hogup Mountain Locality

We interpret the marl units as relatively deepwater and offshore deposits unless there is indication of lagoonal sedimentation patterns and the sand and gravel units as near shore (foreshore/backshore) deposits. The basal gravel (BG) represents a transgressive beach ridge corresponding to a still water level greater than the altitude of the BG and LM interface (1,468 masl). The on lapping lower marls suggest that the gravel marl interface is the backshore face of a truncated beach ridge. Because the crest of this beach ridge has been truncated, it is impossible to accurately estimate the maximum corresponding lake stage altitude that created this ridge from these deposits. However, based on the altitudes and stratigraphic relationships of other local Intermediate beach ridges (Nelson 2012), the altitude of the water level at the time of the BG deposition is estimated to have been between 1,470 and 1,475 masl (Fig. 5.4). The 18,510 ^{14}C yr B.P. age for the BG–LM interface is consistent with the age of other deposits near this altitude found in similar depositional environments (Fig. 5.2; Table 5.1).

The sedimentation patterns of the lower marl (LM) represent a gradual transgression of the lake's water level. Specifically, the abundance of rip up clasts and sand lenses near the base of the deposit are interpreted to represent large storm events in a rising lake level. The sand lenses at the base of the deposit are interpreted as wash over sands that washed down the slope during storms; therefore, as the lake level rose, these sand lenses became thinner as the sand source became more distal. The clay rich marls represent the deepest water (pelagic) deposits preserved in the stratigraphy of the profile. The upper limit of the water level at the time of these clay rich marls cannot be delineated from the sediments of the profile; however, these sediments are estimated to be deposited at a >20 m depth below the water surface, and the lake level is hypothesized to have been near an altitude of 1482 masl (Scott, 1983).

The disconformity at the LG–LM interface is interpreted as evidence of a significant regression in the lake's water level. The undulating and truncated marl surface is evidence of either wave action or subaerial exposure. Based on this interpretation, the lowest limit of the water level during the regression was either near or below the altitude of the LG–LM interface (1469 masl). The 16,480 ^{14}C yr B.P. age for this event does not correlate with other samples in the profile. This anomalous age may be due to multiple contamination sources (meteoric water pooling at marl interface or modern organic material from nearby rootlets). The lower gravels (LG) represent near shore depositional beach gravels (retrograding ridges) that formed as the lake level transgressed upward from the oscillatory event. The 18,240 ^{14}C yr B.P. age of gastropods from the sands at the top of the LG and at the LG–MM (middle marl) interface represent offshore sands as the lake level continued to transgress past the altitude of the LG–MM interface (~1478 msl).

The sandy and silty marls of the middle marl (MM) represent the offshore deposits as the lake level rose. The truncation of these sediments (disconformity) is interpreted as another regression of the lake's water level. The upper limit of the transgression of the lake level prior to this regression is unknown at this locality. The lower limit of the lake dropped to near or below the MM–MG interface (1479–1480 masl). No dateable material was found in the coarse sediment found at this interface. The upper and lower ^{14}C dates below and above the middle gravel unit suggest that the oscillation was between 18,000 and 17,000 ^{14}C yr B.P. The middle near shore gravel represents retro gradational relict beach ridges that were deposited as the lake transgressed past the MM–MG interface.

The transition from the middle gravel (MG) to the upper marl (UM) is interpreted as evidence of another transgression of the lake. The UM is a sandy marl unit representing offshore deposits for which the altitude of the water during deposition is unknown. The truncation of both the MG and UM units overlain by deposits from the upper gravel (UG) is interpreted as the lower limit of a regressive drop in the water level of the lake. The high abundance of gastropods and broken ostracode shells is interpreted as a flushing of older sediments and fossils as the lake rapidly regressed. A 16,930 ± 60 ^{14}C yr B.P. age for the lower MG–UM interface and a 17,210 ± 70 ^{14}C yr B.P. age obtained from the UM–UG interface are within analytical error of each other. We interpret the sediments as the lower limit of a short lived oscillation water level that was near the altitude of the UM–UG interface (~1486 masl) and occurred ~17,000 ^{14}C yr B.P. The deposits of the incised gullies do not suggest an upper limit of the oscillatory event; however, two ages from samples associated with two beach berms to the south (~1.25 km) may be correlated to the active shore face for these lower offshore deposits. A gastropod (sample 25, Table 5.1) from the crest of a gravel berm has an age of 16,930 ^{14}C yr B.P. and an altitude of 1494 masl. Another gravel (younger) berm overlaps the lower (older) berm and has a maximum crest height of ~1512 masl. The crest of the berms was probably at the maximum storm wave base of the deposits, and the crest does not represent the mean water level of the lake at the time of deposition. A gastropod (sample 24, Table 5.1) from fine lagoonal sand in the backshore environment of this younger berm has an age of 16,770 ^{14}C yr B.P. Based on the age and altitudinal relationships, it is inferred that the lower (older) gravel berm is the upper limit of the oscillation, whereas the second (younger) gravel berm is evidence of another transgression of the lake's water level. The sediments of the UM cannot be directly correlated to the lower berm (sample 25); however, if this interpretation is correct, the amplitude of this upper oscillation would be ≥15 m.

Regional Evidence of Intermediate Oscillatory Events

There is little published evidence for oscillations during the Intermediate period of the Lake Bonneville because sedimentological evidence for these oscillations is not well preserved. The few locations documenting these oscillations are fine grained deltaic settings (Sevier Delta—Oviatt, 1997; Bear River Delta—Anderson and Link, 1998) or lagoonal deposits trapped behind beach ridges (Nishizawa, 2010).

Oviatt (1997) suggests that gravel wedges and lag deposits within the fine grained marls and clays of the Sevier Delta near Leamington, Utah (Gilbert, 1890; Varnes and Van Horn, 1991; Oviatt et al., 1994) record oscillations during the transgressive rise of the lake. Oviatt named these oscillations, U1–U3, and suggested that the oscillations occurred at ~18,500–19,000, 17,000–17,500, and 15,000–16,000 ^{14}C yr B.P., respectively (Fig. 5.2) and that the oscillations may have vertical amplitudes similar to the Stansbury oscillation (45 m) (Oviatt, 1997). In addition to Oviatt's sedimentological evidence, $\delta^{18}O$ patterns seen in multiple sediment cores were interpreted as geochemical evidence for these oscillations (Oviatt, 1997). However, recent sediment cores in the northern Bonneville basin (Oviatt, verbal communication, 2011; Benson et al., 2011) do not show the same $\delta^{18}O$ patterns. Therefore, the discordant oscillatory $\delta^{18}O$ patterns of deep water sediment cores may only reflect local changes in the lake chemistry and cannot be used as direct evidence for these smaller oscillatory events.

Anderson and Link (1998) also describe evidence for three large oscillations and multiple (~21) smaller lake fluctuations within the fine grained deltaic deposits of the Bear River. Anderson and Link interpret the sedimentological pattern of the fine grained deposits as evidence of oscillatory events formed through allocyclic processes (i.e., climate change) that may be related to Oviatt's (1997) U1–U3 oscillations. However, no absolute ages have been obtained for these deposits to test the hypothesis. In addition, Anderson and Link's (1998) analysis does not agree with the analysis of Lemons and Chan (1999), who found evidence for only one large oscillation, in the region, and did not describe evidence for small lake level fluctuations. Without good age relationships, it is impossible to relate the deltaic deposits of Anderson and Link (1998) to the proposed oscillations of Oviatt (1997) or the oscillations at the Hogup Mountain locality.

Transgressive deposits within the basin are principally derived from the reworking of Tertiary and Early Pleistocene alluvial deposits or from the excavation of bedrock outcrops. Due to the nature of most sediment sources, transgressive shore faces tend to be less mature (i.e., more angular, poorly sorted, coarser grained) clastic deposits. The lower altitude deposits of the Intermediate paleoshorelines tend to be more mature (i.e., more rounded, well-sorted, finer grained), whereas higher altitude deposits tend to be less mature. This trend is probably due to the low maturity of initial sediment sources and the relatively short duration of the lake at these upper levels. For example, deposits associated with the longer Provo age occupation are more mature than Intermediate and Bonneville sediments because the lacustrine coastal processes reworked sediments deposited during the transgressive rise of the lake into Provo age landforms (Sack, 1999; Godsey et al., 2011). Since much of the fine grained sediments (i.e., offshore marls and sands) deposited in the altitudinal range of the Intermediate paleoshorelines were eroded away and deposited as sediments related to the Provo shore face, complex stratigraphic relationships such as alternating marls and gravels indicating oscillatory patterns are not well preserved. The sediments remaining within the Intermediate altitude range are typically fine grained (marls and sands) related to sediment trapping behind beach ridges, that were buried by coarser sediments deposited during an oscillatory or progradational event of the lake's water level, or were buried by late Pleistocene to Holocene alluvial deposits.

Within the Hogup area and other localities around the basin, there are many coarse Intermediate landforms at multiple altitudes between the Provo and Bonneville paleoshorelines. However, delineating their chronology is difficult because the altitudes of the landform crests are difficult to correlate. The altitudes of these paleoshorelines can differ due to hydro isostatic changes in the basin, changes in wave energy and the resultant wave run up of sediments, varying sedimentation sources and supply, differences in accommodation space, and shorezone slope variations. As mentioned, not many sediments contain dateable radiocarbon material due to the lack of well preserved, fine grained sediments in Intermediate shore faces. Prior to this study, more ages of sediments had been obtained for near shore deposits of the older Intermediate paleoshorelines at lower altitudes (~1,400–1,460 meters above sea level (masl)) than for near shore sediments between 17,500–15,700 ^{14}C yr B.P. and altitudinal ranges of ~1,450–1,524 masl (Table 5.1, Fig. 5.2).

The upper level of the suggested isostatically corrected altitudinal range for Oviatt's (1997) U1 oscillation is ~25 m below the LM/LG interface of the Hogup locality; therefore, evidence for the proposed U1 oscillation is not preserved at the Hogup site. The sedimentological interpretation of the lower oscillation at the Hogup Mountain locality represents a previously undocumented oscillation during the rise of the lake.

The 17,000 to 17,500 ^{14}C yr B.P. age for the proposed U2 oscillation is within the range of altitude of the middle and upper oscillations of the Hogup Mountain locality. The extent of the truncation at the MM/LG interface and the fine grained composition of the offshore marls suggest that the middle oscillation is the greater amplitude of the two proposed oscillations in the Hogup Mountain. Based on the proposed amplitude (~45 m) of Oviatt's (1997) U2, the middle oscillation at Hogup Mountain is most likely related to this oscillation. The upper oscillation in the Hogup Mountain locality is evidence for another undocumented oscillation not correlated to Oviatt's proposed U1–U3 oscillations.

In summary, three potential oscillations are proposed of the altitude range of 1,460–1,512 masl during the transgressive rise of Lake Bonneville. These oscillations are based on the stratigraphical and age control within the Hogup Mountain locality and is the best known example within the basin of a continuous stratigraphic setting that exhibits similar oscillatory events. The lowest of these oscillations occurred between 18,200 and 18,500 ^{14}C yr B.P; the middle oscillation's age is less constrained but occurred between 16,900 and 18,200 ^{14}C yr B.P; and the upper oscillation occurred ~17,000 ^{14}C yr B.P. The amplitude of the proposed Lower Hogup Oscillation (LHO) was ~35–45 m; the middle oscillatory event (correlative to the U2 oscillatory event) had an amplitude of ~45 m (Oviatt, 1997); and the Upper Hogup Oscillation (UHO) had an amplitude of >15 m. A hypothetical adjusted hydrograph of these events is presented in Figure 5.2 and 5.6.

Potential Climatic Teleconnections

Many researchers have proposed global climatic teleconnections in the Great Basin (Zic et al., 2002; Benson et al., 2003; Broecker et al., 2009) and specifically the Bonneville basin (Oviatt, 1997; Oviatt et al., 2005; Broecker et al., 2009; Benson et al., 2011). For example, the H2 Heinrich event is believed to correlate with the timing of the Stansbury Oscillation (Oviatt, 1997, Benson et al., 2011); the cold period of the Younger Dryas stadial (H0) may correlate with the Gilbert level (Oviatt et al., 2005); and the warm period of the Bøelling/Allreød interstadial may be connected to the quick regression from the Provo level (Benson et al., 2011, Godsey et al., 2011) (Fig. 5.6).

The rise of many of the Great Basin lakes, including Lake Bonneville, may be related to the position of the Polar Jet Stream (PJS) during the late Pleistocene (Hostetler and Benson, 1990; Garcia and Stokes, 2006; Benson et al., 2011). This hypothesis states that the Laurentide Ice Sheet (LIS) caused a permanent high pressure system to form above the ice sheet, splitting the PJS into a northern and southern arm and forcing the PJS to the south (Hostetler et al., 1994, Bromwich et al., 2004). When the PJS moved over the lake basin, the regional climate changes by increasing the cloud cover, reducing temperatures, decreasing evaporation, and increasing precipitation (Kutzbach and Guetter, 1986; Hostetler and Benson, 1990; Benson et al., 2011) resulting in an increase of lake size and glacier extent (Benson and Thompson, 1987). As the extent of the LIS fluctuated, resulting in significant ice rafting events (i.e., Heinrich events), the position of the PJS may have also shifted. Smaller lake systems in the Great Basin may have responded quickly with shifts in the PJS (Garcia and Stokes, et al., 2006). However, due to the substantial surface area and volume of Lake Bonneville (Benson et al., 1990; Licciardi, 2001) and the positive feedback of the local glaciers and lake recharge (Munroe et al., 2006; Laabs et al., 2009; 2012), the lake took longer to respond to shifts in the PLS.

Oviatt (1997) proposed that the timing of the Stansbury U1, U2, and U3 oscillations might be correlated to the timing of Heinrich events and/or smaller 1,500 yr cycles (Bond and Lotti, 1995; Stoner et al., 2000). The correlation of basinwide sedimentation patterns, such as unconformable gravel wedges seen in Oviatt (1997) and this study are a better indicator of basinwide oscillatory events than geochemical indicators, such as the deep water marl sediments studied by Oviatt (1997) and Benson et al. (2011). However, the uncertainty of the timing of global climatic events and the proposed oscillatory events makes the correlation of these events somewhat uncertain (Fig. 5.6). Researchers have also suggested that the variation in the highstands of many Great Basin lakes may be caused by regional climatic effects (Allen and Anderson, 1993; Licciardi, 2001; Godsey et al., 2011) so regional climatic variations could also account for oscillatory events such as those described here.

Conclusions

This study strengthens one of Gilbert's (1890) hypotheses that the variation and preservation patterns of the Intermediate paleoshorelines resulted from an oscillating lake during the transgressive phase of Lake Bonneville. This study documents three relatively sizeable oscillations (two that were previously unknown) in the sediments of Hogup Mountain related to Lake Bonneville's transgressive phase. As a result of prior published work and this investigation, seven (7) oscillations are proposed during the transgressive rise of the lake (Fig. 5.6). The proposed oscillations are (from oldest to youngest): two separate oscillations during the Stansbury phase of the lake (Oviatt, 1987; Oviatt et al., 1990; Patrickson et al., 2010); an oscillatory event that Oviatt (1997) calls the U1 event; the lower Hogup oscillatory (LHO) event; an oscillatory event that Oviatt (1997) calls the U2 even; a event we term the upper Hogup oscillatory (UHO) event; and the youngest event that Oviatt (1997) refers to as the U3 event.

The actual number, amplitude, and timing of the oscillations during the Intermediate period of the lake are still relatively uncertain due to the limited stratigraphic and radiometric evidence. Once additional evidence of these oscillatory events are found elsewhere in the basin and the stratigraphic and/or geomorphic evidence for these oscillations better understood, the chronology of the Intermediate paleoshorelines may provide a better representation of the sub millennial changes in the lake's water level. Until that time, it is imperative that further evidence be gathered to better understand the relationships of these oscillatory patterns within the Great Basin and how these patterns may be related to regional and global climatic shifts.

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Figure 5.1: A. Regional map of the extent of Lake Bonneville at its maximum (Bonneville) level in relation to the maximum of other pluvial lakes in the Great Basin; B. The extent of the modern Great Salt Lake in relation to the extent of the Bonneville level and the relation of where the Hogup Mountains and other radiocarbon samples were collected (abbreviations of specific sample locations are listed in Table 1). Other specific localities discussed in the text include Dove Creek (DVC), Hogup Mountains (HM), Zenda Threshold (ZT), Old River Bed (ORB), and Wah Wah Valley (WW).

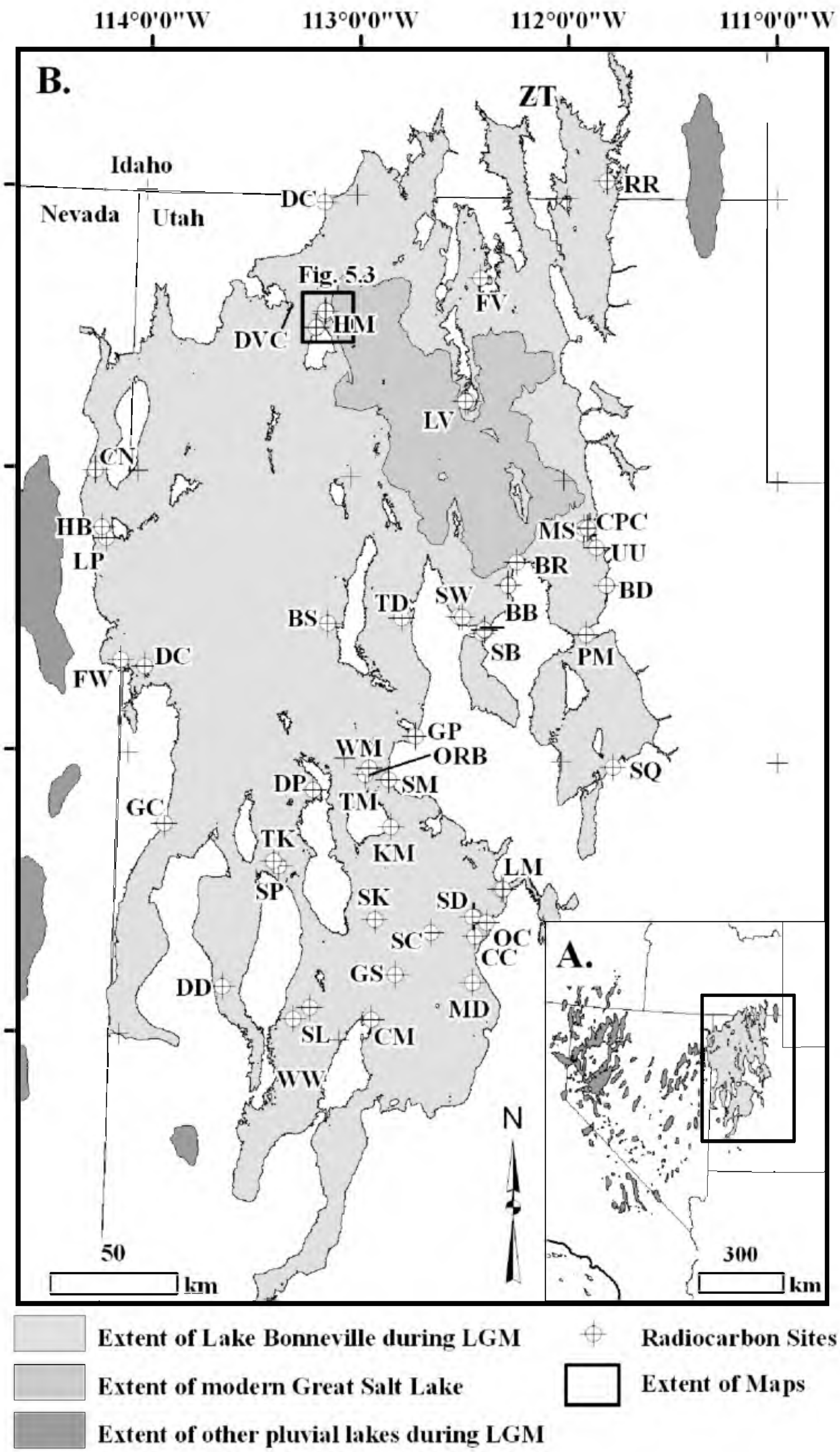
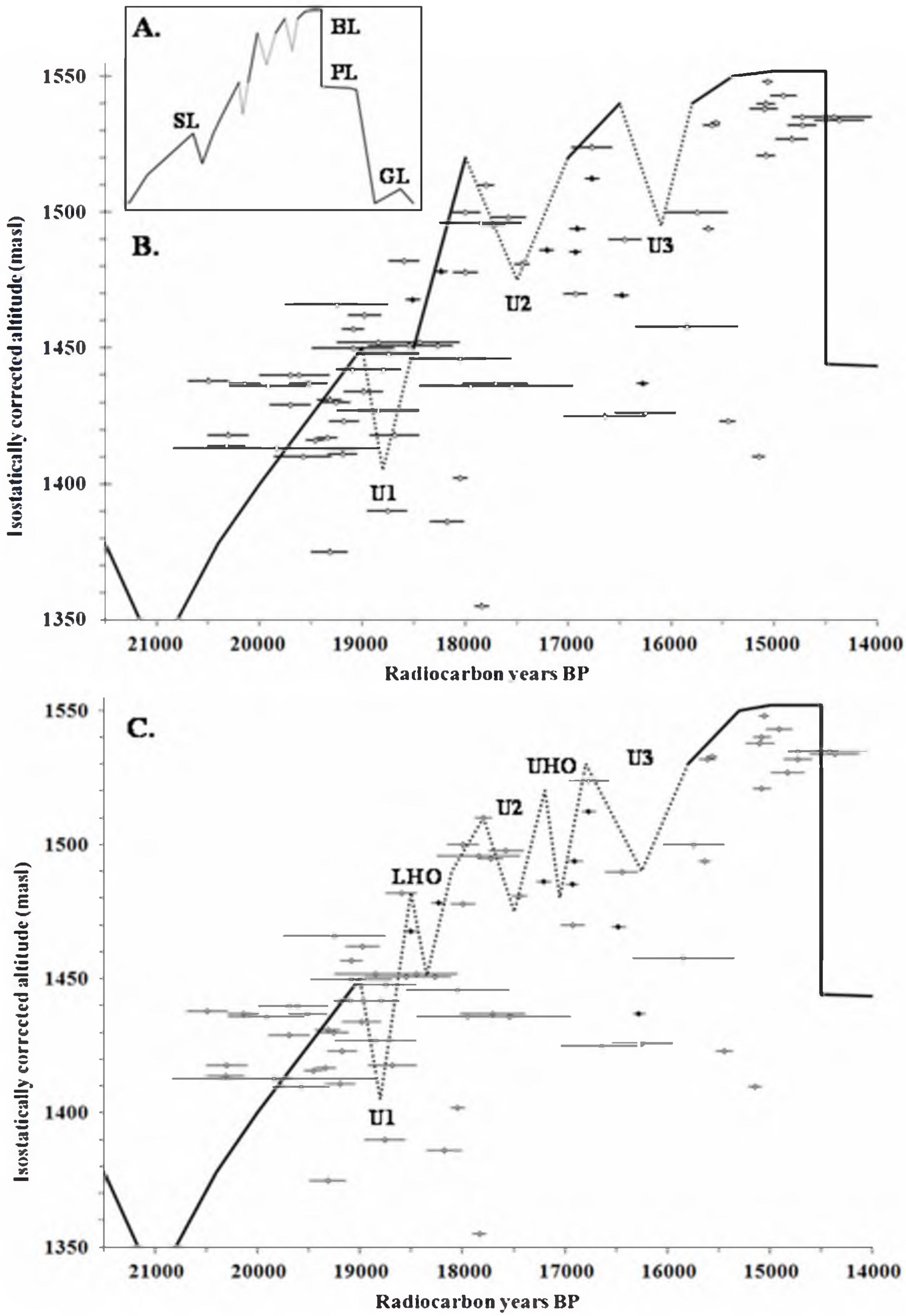

114°0'0"W
113°0'0"W
112°0'0"W
111°0'0"W
42°0'0"N
41°0'0"N
40°0'0"N
39°0'0"N
B.
ZT
Idaho
Nevada
Utah
DC
RR
Fig. 5.3
FV
HM
DVC
LV
CN
HB
LP
MS
CPC
UU
BR
BD
TD
SW
BB
BS
SB
DC
FW
PM
GP
WM
ORB
DP
SM
TM
GC
SQ
TK
KM
LM
SP
SK
SD
SC
OC
CC
GS
DD
MD
SL
CM
WW
N
50
km
A.
300
km
Extent of Lake Bonneville during LGM
Extent of modern Great Salt Lake
Extent of other pluvial lakes during LGM
Radiocarbon Sites
Extent of Maps

Figure 5.2: A: Generalized hydrograph of Lake Bonneville modified from Oviatt (1997) and Godsey et al. (2011). B: Plot of age and isostatically corrected altitudes of samples in Table 1 in relation to the Lake Bonneville hydrograph. C: Modified hydrograph with proposed oscillatory events. Amplitude limits of water level fluctuations associated with the U1, U2, and U3 oscillations are approximate and shown here schematically. Open circles indicate previously published data, whereas closed circles represent data from this study. The horizontal lines emanating from the sample points represent the error bars related to the age of the sample.


A.
BL
PL
SL
GL
B.
1550
1500
1450
1400
1350
Isostatically corrected altitude (masl)
U1
U2
U3
21000
20000
19000
18000
17000
16000
15000
14000
Radiocarbon years BP
C.
LHO
UHO
U2
U3
U1

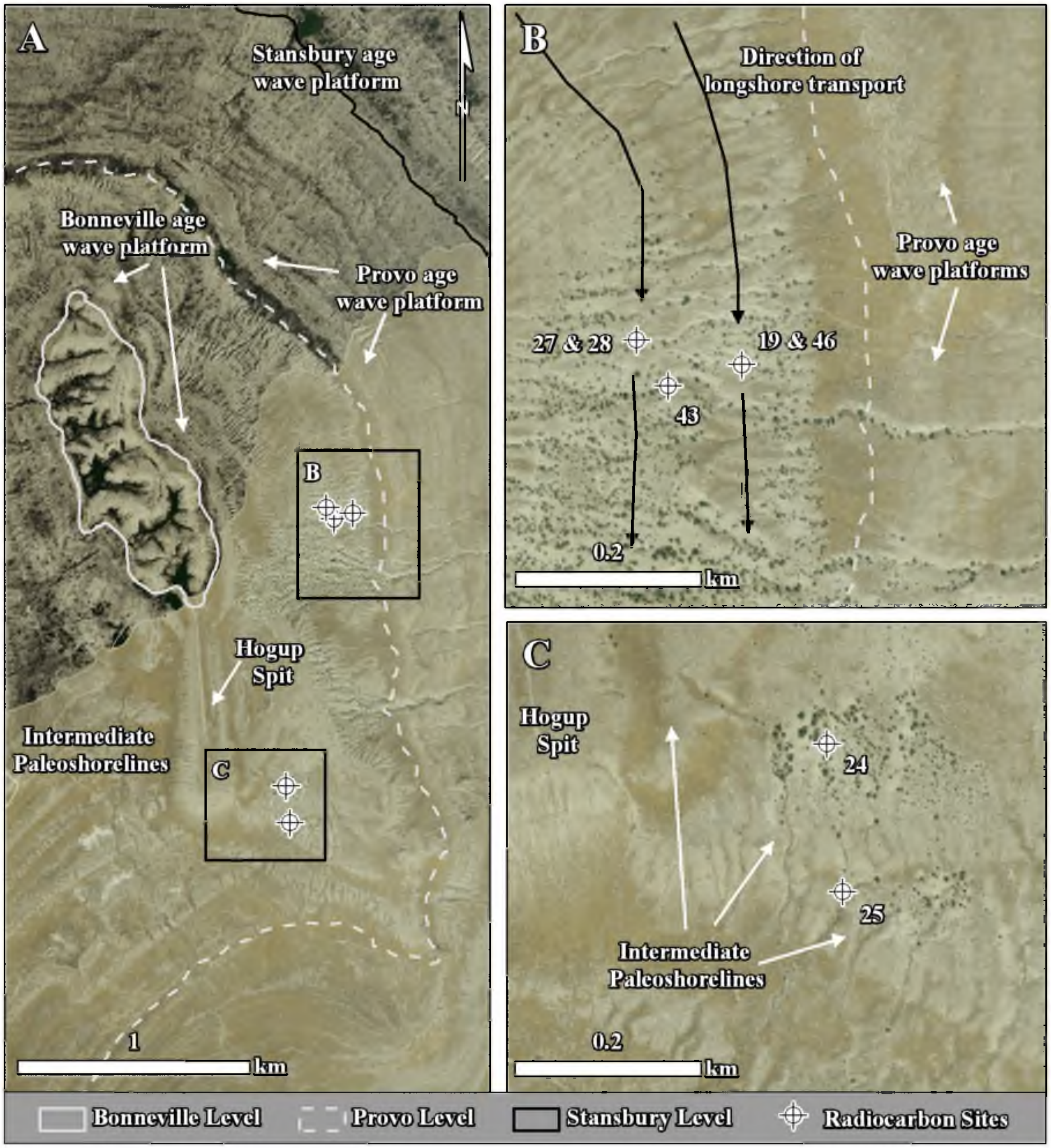
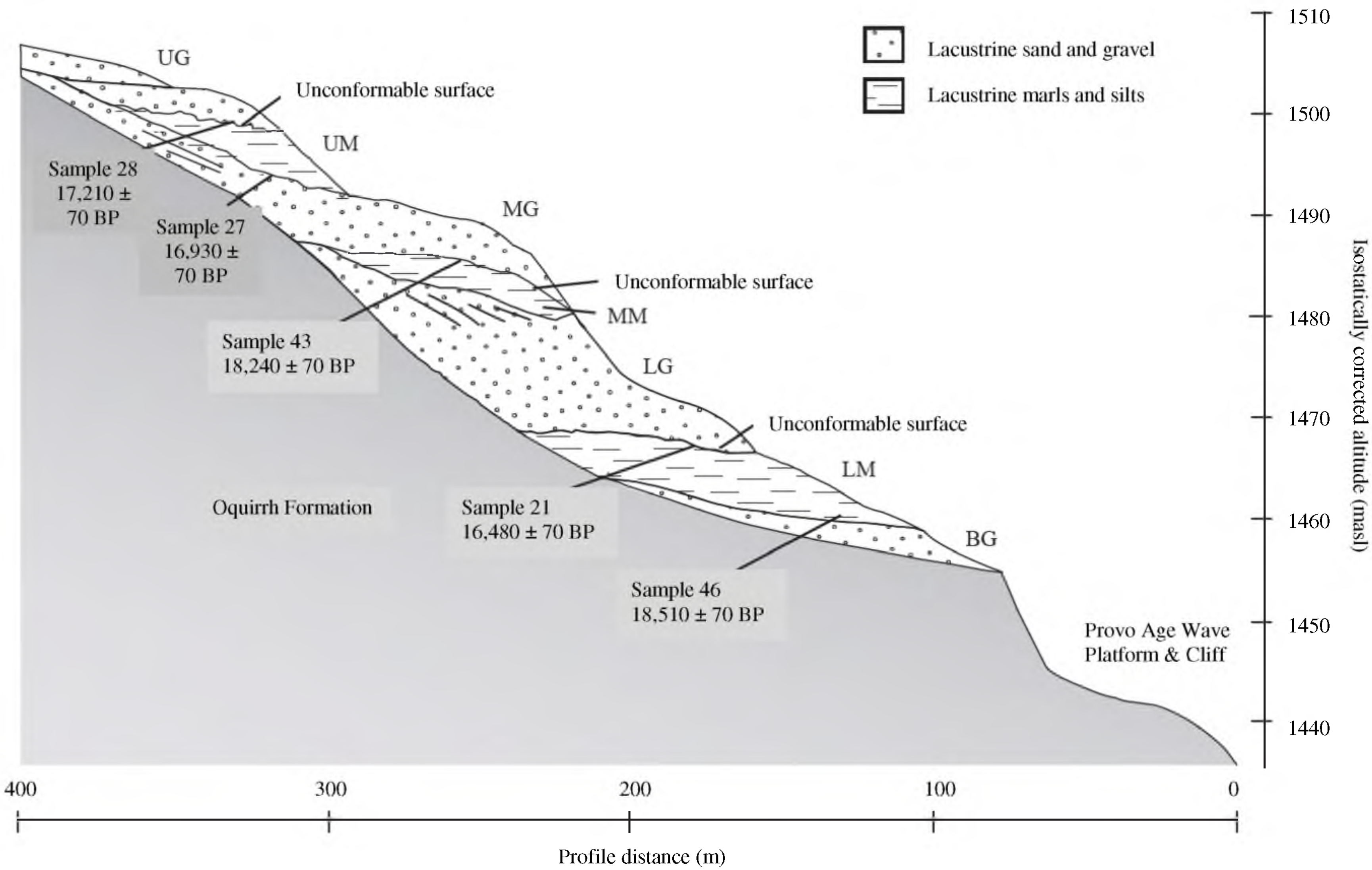


Figure 5.3: Orthophoto images of the Hogup Mountain area. Sample locality information can be found in Table 1, and location of area in reference to the entire Bonneville basin can be seen in Figure 1. A. Overview of the Hogup Mountain area (Black boxes indicate the extent of A and B); B. Close up of where the sections for the composite profile in Figure 4 were measured and where radiocarbon samples were collected; C. Close up of beach ridges where radiocarbon samples 24 and 25 were collected.

Figure 5.4: Schematic composite profile of lacustrine units associated with the Intermediate deposits of the Bonneville Lake Cycle within the Hogup Mountain area. Abbreviations of the units are as follows: BG—basal gravel, LM—lower marl, LG—lower gravel, MM—middle marl, MG—middle gravel, UM—upper marl, and UG—upper gravel.



Lacustrine sand and gravel
Lacustrine marls and silts
UG
Unconformable surface
UM
Sample 28
17,210 ±
70 BP
Sample 27
16,930 ±
70 BP
MG
Unconformable surface
MM
Sample 43
18,240 ± 70 BP
LG
Unconformable surface
LM
Oquirrh Formation
Sample 21
16,480 ± 70 BP
BG
Sample 46
18,510 ± 70 BP
Provo Age Wave
Platform & Cliff
1510
1500
1490
1480
1470
1460
1450
1440
Isostatically corrected altitude (masl)
400
300
200
100
0
Profile distance (m)

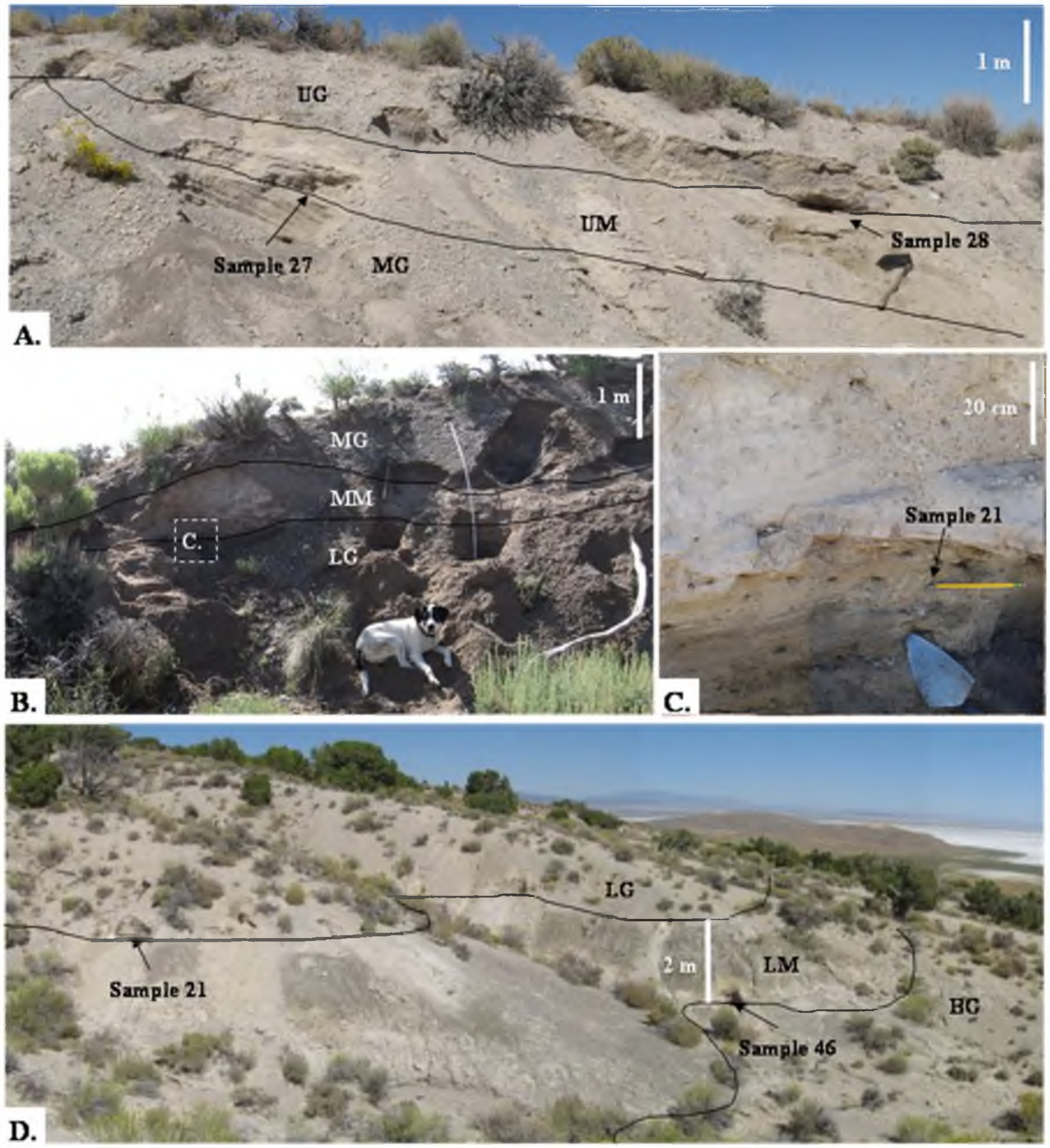


Figure 5.5: Field photographs from the Hogup Mountain composite profile: A. Upper marl (UM) pinching out between the middle gravel (MG) and upper gravel (UG); B. Middle marl (MM) truncated and overlain by MG; C. Ripple laminated sands overlain by sandy gravel; D. Basal gravels (BG) conformably overlain by lower marls (LM) and then unconformably overlain by lower gravels (LG).

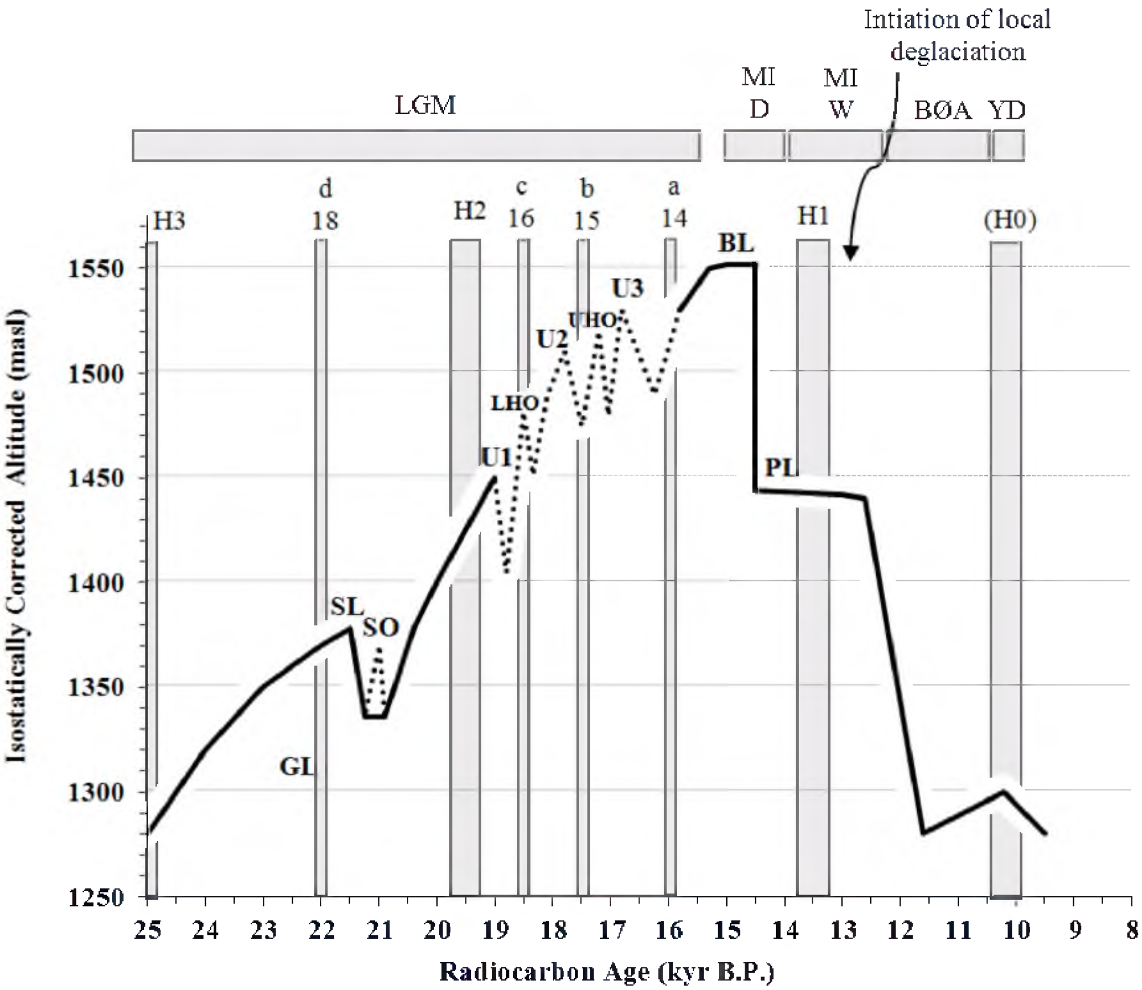


Figure 5.6: Hypothesized hydrograph modified from Oviatt (1997) and Godsey et al. (2011) compared with potential global climatic teleconnections. Altitudes are adjusted for effects of differential isostatic rebound in the basin. Amplitude limits of water level fluctuations associated with the Intermediate oscillations are approximate and shown here schematically. The age of the local deglaciation is based on ^{10}Be exposure ages for boulders within moraines within the local Wasatch and Unita Ranges (Refsnider et al., 2008; Laabs et al., 2011). The timing of North Atlantic Heinrich events (H0–H3) and 1,500 yr climatic events (a/14, b/15, c/16, and d/18) are based on ages from Stoner et al. (2000) and Oviatt's (1997) interpretation of Bond and Lotti (1995). Abbreviations are as follows; S—Stansbury level, SO—Stansbury oscillations, Intermediate Oscillations (U1, LHO, U2/UHO, and U3), B—Bonneville level, P—Provo level, G—Gilbert level, YD—Younger Dryas stadial, BØA—the Bølling-Allerød interstadial, MI—Mystery Interval (W—wet and D—dry) (Broecker et al., 2009; Denton et al., 2006); LGM—Last Glacial Maximum (Clark, 2009).

Table 5.1: Radiocarbon Ages. Radiocarbon data used in this report including a brief description of the stratigraphic context of the sample, the measured and isostatically corrected altitude, and the location as seen in Figure 5.1. Bon. (Bonneville Paleoshoreline).

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material	Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
1	BR	Beta-5697	40.708	112.215	1,570 (1,534)	rad	Stagnicola	14,370 ± 240	18,037 ---------- 16,887	Shore face sand	Godsey et al., 2005
2	SB	Beta-50770	40.466	112.363	1,576 (1,535)	rad	Stagnicola	14,420 ± 370	18,503 - 18219 ---------- 18,194 – 16,856	From interstices of tufa on gravel beach crest below the Bon.	Godsey et al., 2005
3	SB	SI-4277C	40.466	112.363	1,576 (1,535)	rad	lacustrine tufa TIC	14,730 ± 100	18,496 – 18,259 ---------- 18,109 – 17,579	Tufa on gravel beach ridge crest below the Bonn.	Benson et al., 1990
4	SB	Beta-146004	40.465	112.360	1,572 (1,532)	rad	Stagnicola	14,730 ± 140	18,512 – 18,205 ---------- 18189 - 17530	Laminated med - fine sand overlain by coarse sand & gravel in embayment, ~ 30 m below the Bon.	Godsey et al., 2005
5	SM	Beta-39294	39.928	112.796	1,560 (1,527)	rad	Stagnicola	14,830 ± 160	18,538 ---------- 17,644	Lacustrine sand 1 m above boulder line	Oviatt et al., 1994
6	DCH	Beta-29019	41.973	113.154	1,566 (1,543)	rad	Pyrgulopsis	14,910 ± 130	18,554 ---------- 17,794	Backshore gravelly sand exposed in highway cut	Godsey et al., 2005
7	SW	Beta-169099	40.510	112.467	1,594 (1,548)	AMS	Stagnicola	15,060 ± 50	18,561 – 18,406 ---------- 18,393 – 18,026	Marly mud on the landward side of spit ~ 6 m below Bon.	Godsey et al., 2005
8	GC	Beta-156852	39.753	113.823	1,558 (1,521)	rad	Stagnicola	15,080 ± 90	18,581 ---------- 18,018	Laminated gravely sand just below the Bon.	Godsey et al., 2005

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
9	SW	Beta-151451	40.512	112.483	1,594 (1,540)	rad	Stagnicola
10	CPC	W-5261	40.832	111.888	1,570 (1,538)	rad	Wood
11	DM	CURL-6018	39.413	112.452	1,425 (1,410)	AMS	Pyrgulopsis
12	LM	L-774N	39.546	112.259	1,526 (1,500)	rad	gastropod shells
13	SK	CURL-6017	39.430	112.845	1,440 (1,423)	AMS	Stagnicola
14	PM	Beta-168998	40.462	111.902	1,564 (1,533)	AMS	Pyrgulopsis
15	GCP	Beta-156672	40.083	112.675	1,562 (1,532)	rad	Stagnicola
16	SG	AA-19048	39.983	111.759	1,515 (1,494)	AMS	Pyrgulopsis

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
15,080 ± 90	18,581 ----------- 18,018	Rippled laminated gravelly sand on basinward side of spit ~ 20 m below the Bon.	Godsey et al., 2005
15,100 ± 140	18,633 ----------- 17,978	Organic mud, behind a buried transgressive lagoon/bar complex	Scott, 1988
15,150 ± 65	18,621 – 18,435 ----------- 18,366 – 18,033	Delta front	Kaufman, 2003
15,400 ± 300	19,317 ----------- 17,958	Basal 30 cm white calcareous sand right above thin gravel wedge	Broecker & Kaufman, 1965
15,450 ± 80	18,846 ----------- 18,541	---------------------	Kaufman, 2003
15,570 ± 50	18,887 ----------- 18,615	30 cm thinly bedded laminated coarse silt overlying a lagoonal fine sand wash over deposit	Nishizawa, 2010
15,610 ± 80	18,933 ----------- 18,607	Transgressive-phase facies on the wash over side of a gravel beach barrier. Sample taken from top of silty gravel (4 m) that is overlain by 2 m thick coarse silt.	Nishizawa-10
15,640 ± 55	18,977 ----------- 18,594	Sand from backshore face	Light, 1996

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
17	OC	L-774H	39.428	112.333	1,481 (1,458)	rad	gastropod shells
18	LV	L-775I	41.277	112.470	1,460 (1,426)	rad	gastropod shells
19	HG	Beta-246725	41.507	113.217	1,470 (1,437)	AMS	Stagnicola
20	SB	Beta-10245	40.453	112.374	1,506 (1,490)	rad	gastropod shells
21	HG	Beta-307249	41.589	113.139	1,507 (1,470)	AMS	Stagnicola
17	OC	L-774H	39.428	112.333	1,481 (1,458)	rad	gastropod shells
18	LV	L-775I	41.277	112.470	1,460 (1,426)	rad	gastropod shells
19	HG	Beta-246725	41.507	113.217	1,470 (1,437)	AMS	Stagnicola

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
15,850 ± 500	20,123 ----------- 17,999	Collected from marl 8 - 30 cm above Pahvant Butte (basaltic) ash	Broecker & Kaufman , 1965
16,250 ± 300	20,113 ----------- 18,806	Sand lens separating two sequences of white marl	Broecker & Kaufman , 1965
16,280 ± 60	21,479 (21,409) 21,229	Gravel barrier - reverse grading deposit from coarse gravels and sand overlying finer sands	this study
16,450 ± 160	20,134 ----------- 19,297	Gravel barrier	Burr & Currey, 1992
16,480 ± 70	21,759 – 21,539 (21,499) 21,649 – 21,389	Fine sand above clay/silt rich marls underlying beach gravels	this study
15,850 ± 500	20,123 ----------- 17,999	Collected from marl 8 - 30 cm above Pahvant Butte (basaltic) ash	Broecker & Kaufman , 1965
16,250 ± 300	20,113 ----------- 18,806	Sand lens separating two sequences of white marl	Broecker & Kaufman , 1965
16,280 ± 60	21,479 (21,409) 21,229	Gravel barrier - reverse grading deposit from coarse gravels and sand overlying finer sands	this study

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
20	SB	Beta-10245	40.453	112.374	1,506 (1,490)	rad	gastropod shells
21	HG	Beta-307249	41.589	113.139	1,507 (1,470)	AMS	Stagnicola
22	LV	L-775J	41.277	112.470	1,459 (1,425)	rad	gastropod shells
23	CPC	W-4896	40.453	111.888	1,553 (1,523)	rad	wood
24	HG	Beta-246723	41.578	113.141	1,556 (1,513)	AMS	Stagnicola
25	HG	Beta-246724	41.577	113.141	1,512 (1,474)	AMS	Stagnicola
26	LV	AA-19042	41.276	112.455	1,518 (1,470)	AMS	Pyrgulopsis
27	HG	Beta-307250	41.588	113.138	1,525 (1,485)	AMS	Stagnicola

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
16,450 ± 160	20,134 ----------- 19,297	Gravel barrier	Burr & Currey, 1992
16,480 ± 70	21,759 – 21,539 (21,499) 21,649 – 21,389	Fine sand above clay/silt rich marls underlying beach gravels	this study
16,650 ± 400	20,957 – 20,672 ----------- 20,576 – 18,887	Sandy beach foreshore toe	Broecker & Kaufman, 1965
16,770 ± 200	20,332 ----------- 19,468	Buried transgressive lagoon/bar complex	Scott, 1983
16,770 ± 70	22,069 – 21,749 (21,859) 21,589 – 21,579	Fine sand lagoon deposits behind intermediate gravel barrier	this study
16,910 ± 80	22,229 (21,979) 21,809	Within slightly cemented intermediate gravel barrier	this study
16,930 ± 120	20,396 – 19,800 ----------- 19,697 – 19,591	Sandy beach foreshore toe	Light, 1996
16,930 ± 60	22,219 (22,119) 21,989	Fine sand above beach gravels that gradates into a sandy marl	this study

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
28	HG	Beta-307251	41.589	113.139	1,526 (1,486)	AMS	Stagnicola
29	CM	CURL-6014	39.075	112.852	1,505 (1,481)	AMS	Stagnicola
30	LM	L-711C	39.546	112.268	1,506 (1,496)	rad	gastropod shells
31	SRD	L-774A	39.446	112.396	1,453 (1,436)	rad	Stagnicola
32	BCC	W-4451	40.630	111.795	1,524 (1,498)	rad	charcoal
33	DCD	Beta-26795	39.181	113.538	1,450 (1,437)	rad	gastropod shells
34	BCB	Beta-153159	40.624	112.254	1,520 (1,495)	rad	Stagnicola

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
17,210 ± 70	22,439 (22,319) 22,239	Fine sand above silt/sand rich marls underlying beach gravels	this study
17,450 ± 75	21,226 ----------- 20,401	Sandy and gravelly backshore face (?)	Kaufman , 2003
17,500 ± 400	21,920 – 19,851 ----------- 19,666 – 19,717	Top of coarse gravel wedge between marls	Broecker & Kaufman , 1965
17,550 ± 600	22,301 ----------- 19,536	Silty sand 15 - 30 cm above green calcareous clay - top of delta (?)	Broecker & Kaufman , 1965
17,580 ± 170	21,418 ----------- 20,415	4 cm thick, organic rich silt within 1m thick fine-grained alluvium that overlies a soil	Scott, 1983
17,710 ± 320	22,010 (21,060) 20,287	Shore face gravel (?)	Sack, 1999
17,730 ± 120	21,490 – 20,775 ----------- 20,739 – 20,545	Gravelly sand from backshore face	Nishizawa, 2010

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
35	KM	CURL-6019	39.760	112.780	1,540 (1,510)	AMS	Stagnicola
36	BST	Beta-169219	40.479	113.091	1,365 (1,355)	AMS	Stagnicola
37	SRD	L-672C	39.446	112.393	1,453 (1,436)	rad	ostracodes
38	EB	Beta-16911	40.628	112.256	1,502 (1,478)	rad	wood
39	BCC	W-4687	40.630	111.795	1,524 (1,500)	rad	wood
40	SRD	L-774A	39.446	112.396	1,453 (1,446)	rad	Pyrgulopsis
41	FV	Beta-168083	41.715	112.408	1,420 (1,402)	AMS	Stagnicola

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
17,800 ± 80	21,534 – 20,875 ----------- 20,650 – 20,597	Sandy foreshore face	Kaufman, 2003
17,840 ± 70	21,549 ----------- 20,994	1.7 m medium-fine sand (sampled from bottom 20 cm) overlying a calcareous fine sand followed by a sandy marl	Nishizawa, 2011
17,950 ± 500	22,537 ----------- 20,114	Sand immediately overlying green calcareous clay equivalent to deep white marl - top of delta	Broecker & Kaufman, 1965
18,000 ± 120	22,027 ----------- 21,160	Lagoonal organic mud, behind a buried transgressive lagoon/bar complex	Nishizawa, 2010
18,000 ± 150	22,097 ----------- 21,073	50 cm thick fine-grained alluvium & marsh deposit that is conformably overlain by lake gravel and sand	Scott, 1983
18,050 ± 500	22,872 – 22,848 ----------- 22,657 – 20,176	Silty sand 15 - 30 cm above green calcareous clay - top of delta (?)	Broecker & Kaufman, 1965
18,050 ± 70	21,924 ----------- 21,257	Laminated (4 m) lagoonal fine sand behind a baymouth barrier overlain by >2 m laminated marl. Sample taken from top unit.	Nishizawa, 2010

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
42	WM	Beta-39295	39.968	112.886	1,377 (1,386)	rad	Pyrgulopsis
43	HG	Beta-307247	41.588	113.139	1,517 (1,478.3)	AMS	Stagnicola
44	DP	Beta-153160	39.886	113.141	1,472 (1,451)	rad	Stagnicola
45	MD	L-774I	39.213	112.392	1,458 (1,452)	rad	gastropod shells
46	HG	Beta-307248	41.588	113.138	1,505 (1,467.7)	AMS	Stagnicola
47	SRD	L-774C	39.447	112.387	1,457 (1,451)	rad	gastropod shells
48	PM	W-4693	40.466	111.907	1,493 (1,482)	rad	Picea wood

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
18,180 ± 170	22,235 ----------- 21,329	Sandy beach foreshore toe	Oviatt, 1994
18,240 ± 70	24,009 (23,539) 23,459	Fine sand above beach gravels that gradates into a sandy marl	this study
18,270 ± 150	22,256 ----------- 21,435	Gravelly sand unit on the wash over side of a gravel barrier - the gravelly sand grades upward into a gravel.	Nishizawa, 2010
18,450 ± 400	23,277 – 23,061 ----------- 23,054 – 21,084	15 cm thick sand layer immediately below a deep white marl.	Broecker & Kaufman,
18,510 ± 70	24,239 (24,129) 23,999	Fine sand above beach gravels that gradates into a sandy marl	this study
18,550 ± 400	23,310 ----------- 21,248	Gravelly sand immediately under green calcareous clay equivalent to deep white marl - delta front (?)	Broecker & Kaufman, 1965
18,600 ± 150	22,493 ----------- 21,577	30 cm thick mud that overlies soil developed in sand & gravel of Little Valley lake cycle (?) - mud is overlain by sand & gravel of the Point of the Mountain Spit	Scott, 1983

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
49	DCD	Beta-186151	40.309	113.939	1,438 (1,418)	rad	Stagnicola
50	SRD	L-672A	39.446	112.392	1,454 (1,448)	rad	gastropod shells
51	SL	Beta-36192	39.072	113.208	1,400 (1,390)	rad	Stagnicola
57	TAD	Beta-172988	40.503	112.748	1,491 (1,457)	AMS	Stagnicola
58	CPC	W-5263	40.833	111.904	1,473 (1,450)	rad	wood
59	CNN	Beta-57131	41.000	114.200	(1,442)	rad	mollusk shells
60	HHS	Beta-191601	40.399	111.932	1,440 (1423)	AMS	Stagnicola
61	GSF	AA-19049	39.235	112.748	1,424 (1411)	AMS	Pyrgulopsis

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
18,690 ± 240	23,221 – 23,139 ----------- 22,973 – 21,531	~3 cm gravelly medium sand overlying a gravel spit. The sand is overlain by >60 cm of boulders with dendritic tufa followed by a white marl	Nishizawa, 2010
18,750 ± 300	23,276 ----------- 21,543	Base of green calcareous clay equivalent to deep white marl	Broecker & Kaufman, 1965
18,760 ± 200	23,260 – 23,105 ----------- 22,999 – 21,620	Transgressive-phase shorezone from white sandy marl	Oviatt et al., 1992
19,090 ± 110	23,297 ----------- 22,385	Lagoonal fine sand behind a gravel barrier beach	Nishizawa, 2010
19,090 ± 400	23,811 ----------- 21,752	Buried transgressive lagoon/bar complex	Scott, 1988
19,100 ± 160	23,339 (22,660) 22,352	Shore face gravel (?)	Sack, 1999
19,180 ± 145	23,376 ----------- 22,420	Sand from beach foreshore toe	Nishizawa, 2010
19,195 ± 145	23,386 ----------- 22,430	Marly sand	Light, 1996

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
62	RRP	W-982	42.065	111.811	1,478 (1466)	rad	Stagnicola
63	FW	Beta-47753	40.330	115.050	1,430 (1430)	rad	mollusk shells
64	HB	Beta-156673	40.796	114.159	1,441 (1431)	rad	Stagnicola
65	SP	Beta-27462	39.361	113.318	1,387 (1375)	rad	Pyrgulops is
66	DCD	Beta-186153	40.308	113.934	1,437 (1417)	AMS	Stagnicola
67	DCD	Beta-169981	40.309	113.939	1,434 (1416)	AMS	Pyrgulops is
68	LP	Beta-17622	40.758	114.137	(1,437)	rad	Stagnicola
69	LV	W-4445	41.277	112.470	1,442 (1,410)	rad	wood

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
19,250 ± 500	24,222 ----------- 21,721	Marly lagoonal deposits	Bright, 1963
19,260 ± 140	23,431 (22,840) 22,476	Shore face gravel (?)	Sack, 1999
19,320 ± 120	23,470 ----------- 22,548	Gravelly sand unit on a transgressive bayhead gravel barrier	Nishizawa, 2010
19,320 ± 180	23,549 ----------- 22,461	Sandy marl - base of white marl above transgressive-phase barrier beach, shallow water in offshore deposits	Oviatt, 1991
19,340 ± 100	23,466 ----------- 22,592	Fine sand overlying tufa-capped boulders on a transgressive bayhead gravel barrier	Nishizawa, 2010
19,460 ± 90	23,587 ----------- 22,665	Sandy gravel matrix on the wash over side of a spit - beach backshore face (typically sandy and gravelly)	Nishizawa, 2010
19,520 ± 190	23,824 (23,140) 22,617	Gravelly sand from beach backshore face	Sack, 1999
19,580 ± 280	24,071 ----------- 22,503	Lagoonal deposits within a gravel bar complex	Scott, 1983

Table 5.1: *(continued)* Radiocarbon Ages.

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
70	CPC	W-5326	40.833	111.904	1,463 (1,441)	rad	wood
71	CPC	W-4421	40.833	111.904	1,445 (1,424)	rad	wood
72	CPC	W-5272	40.833	111.904	1,463 (1,441)	rad	wood
73	UU	W-1743	40.762	111.845	(1,413)	rad	tree bark
74	TK	Beta-27463	39.612	113.293	(1,436)	rad	Stagnicola
75	TK	AA-19058	39.612	113.293	(1,437)	AMS	Stagnicola
76	LV	AA-19065	41.278	112.468	(1,418)	AMS	Stagnicola
77	SCB	AA-19051	39.387	112.589	(1,414)	AMS	Sphaerium

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
19,620 ± 300	24,189 ----------- 22,539	Buried transgressive lagoon/bar complex	Scott, 1988
19,700 ± 200	24,136 – 22,917 ----------- 22,810 – 22,683	2.5 m below top of 9 m thick lagoonal deposit that interfingers above and below with gravel bar complexes. Overlies buried soil formed in older lake deposits.	Scott, 1983
19,700 ± 300	22,005 ----------- 24,278	Buried transgressive lagoon/bar complex	Scott, 1988
19,840 ± 1000	26,262 ----------- 21,237	Shore face gravel (?)	Mars et-69
19,920 ± 380	24,571 ----------- 22,628	Calcareous sandy mud ~5.5 m below the Provo, transgressive-phase shallow offshore deposit	Oviatt, 1991
20,145 ± 155	24,491 ----------- 23,623	Beach foreshore toe (typically sand)[d]	Light, 1996
20,305 ± 200	24,845 ----------- 23,731	Beach backshore face (typically sand & gravelly)[d]	Light, 1996
20,320 ± 185	24,826 ----------- 23,781	------------------	Light, 1996

Num	Site	Lab Num.	Lat.	Long.	Altitude[a]	Method	Dated Material
78	PH	Beta-172664	41.564	112.605	(1,428)	AMS	Pyrgulopsis
79	PH	Beta-172665	41.571	112.595	(1,448)	AMS	Stagnicola
80	PH	Beta-168082	41.560	112.607	(1,417)	AMS	Pyrgulops is
81	CPC	SI-4124	40.833	111.904	1,438 (1,417)	rad	wood
82	SV	Beta-52614	40.828	112.798	(1,431)	rad	mollusk shells
83	PH	Beta-172173	41.561	112.607	(1,420)	AMS	Pyrgulops is
84	PH	Beta-172174	41.572	112.594	(1,450)	AMS	Stagnicola
85	FS	Beta-47752	39.608	113.283	(1,433)	rad	mollusk shells
86	JN	W-4897	40.457	111.917	1,405 (1,417)	rad	Abies or Juniperus wood

Radiocarbon age (^{14}C yr B.P.)	Calibrated age 2 sigma (cal BP)[b]	Depositional Setting	Source
20,440 ± 90	24,785 ----------- 23,974	Muddy - very fine sands of the beach foreshore toe (typically sandy) overlying a gravel barrier	Nishizawa, 2010
20,480 ± 90	24,891 ----------- 24,081	Muddy - very fine sands of the beach foreshore toe (typically sandy) overlying a gravel barrier	Nishizawa, 2010
20.490 ± 130	24,920 ----------- 24,012	Muddy - very fine sands of the beach foreshore toe (typically sandy) overlying a gravel barrier	Nishizawa, 2010
20.500 ± 200	22,862 ----------- 24,981	Buried transgressive lagoon/bar complex	Scott, 1988
20.540 ± 280	25,183 ----------- 23,793	Gravel from beach foreshore face[d]	Sack, 1999
20.600 ± 150	25,021 ----------- 24,192	Muddy - very fine sands of the beach foreshore toe (typically sandy) overlying a gravel barrier	Nishizawa, 2010
20.630 ± 150	25,032 ----------- 24,234	Muddy - very fine sands of a beach foreshore face (typically gravelly)	Nishizawa, 2010
20.830 ± 420	25,998 ----------- 23,861	Gravel from beach foreshore face[d]	Sack, 1999
20.900 ± 250	25,704 ----------- 24,316	Middle of 1.3 m thick, brown-dark gray lens of lagoonal mud within gravel & sand of a bar complex	Scott, 1983